

CHAPTER TWO
Sediment Transport

2.1 Introduction

▪ **Need for Understanding of Sediment Transport in Rivers**

Understanding of sediment transport processes is essential for integrated river management and river engineering. A river is not only conveying water, but has many other functions. One of these functions is the transport of erosion products (boulders, gravel, sand silt and clays) from its catchment in the downstream direction. If the transport capacity of the river is affected, e.g. by diversion of water from the river or by storing water in a reservoir, deposition of sediments may occur. If not properly taken care of, harmful sedimentation and/or erosion may occur due to water management measures, which then again to be remedied. Hence, good water management (and more generally good river management, which includes also the management of, e.g. flood plain of the river) includes sediment management. For this, sufficient understanding of the sediment transport processes in the river is a must.

There are, however, other reasons why good understanding of sediment transport process is indispensable. The most important ones are listed hereafter:

(a) Morphological boundary conditions for design of hydraulic structures and river training works.

For the design of such structures boundary conditions have to be specified. These comprise not only the discharge, maximum and minimum water levels and current velocities, but also the lowest bed levels that may occur in a river and the future changes in the morphological characteristics of the river (changes in the “form” of the river; morphology is the science of the form of the river) near the structure. These levels and possible changes are effectuated by gradients in the sediment transport rates. Changes are not only due to natural processes like the variation of the discharge in time or ongoing bank erosion, but there will also be an interaction between the future structure and the nearby flow pattern and sediment transport. Also the latter should be accounted for properly. This requires understanding of the interaction between flow, sediment and structure.

(b) Sedimentation in Reservoirs

Many reservoirs are suffering from excessive sedimentation. Often this is due to the fact that either the upstream sediment supply was never considered or that the seriousness of this process was underestimated, e.g. because not sufficient data were available. Also changes in sediment yield, e.g. due to changed land use in the upstream catchment, can cause detrimental sedimentation. To remedy the negative effects (loss in storage capacity), changes in the operation of the reservoir are often required, sometimes with drastic consequences, e.g. the reduction of the power production.

(c) Sediment problems at Intakes

Many main canals of irrigation projects suffer from excessive sedimentation, which is entering via the headworks. Often this is due to the fact that the sediment transport in the river was not properly assessed and appears to be much larger than anticipated. Another

reason is that morphological changes have taken place after construction of the intake structure, which jeopardized the measures taken to exclude sediment when withdrawing water. Remedial measures are rehabilitation of the intake structure or regular cleaning of the canals, which both are very expensive.

(d) Environmental Impact Assessment

There is an increasing awareness of possible environmental impact of water management schemes and river training works. To assess the potential impacts of such schemes not only their hydraulic impacts should be assessed but also possible morphological changes (degradation or aggradation, changes in flood plain sedimentation, etc) should be identified. This assessment can only be done when sufficient understanding is available on sediment transport processes.

Many more examples can be given which all stress the importance of understanding sediment loads and sediment transport processes in rivers before embarking on major water management schemes or river training.

These lecture notes attempt to provide the student with a basic understanding of sediment transport in rivers. Necessarily it can only give an introduction to sediment transport processes.

2.2 Origin and properties of sediments

2.2.1 Introduction

The processes of erosion of land surfaces, transportation of eroded materials, deposition of this material in lakes and reservoirs and such other processes depend on several factors. These factors can be generally classified into the following categories: characteristics of sediment, characteristics of the fluid, characteristics of the flow, and characteristics of the channel.

The sediment characteristics, such as its relative density, size, shape, etc play a decisive role in various stages of the phenomenon of sediment transport and, therefore, these will be discussed here in detail along with the methods of determining these properties. These properties are governed to a large extent by the origin of the sediment and the process of its formation. Hence attention will first be directed towards the origin and formation of sediments.

2.2.2 Origin and formation of sediments

All sediments transported by water and air and also those found in deserts have resulted by the process of weathering of rocks. Weathering can be defined as the process by which solid rocks are broken up and decayed. The size, the mineral composition, the density and other factors such as surface texture depend on the nature of the parent rock from which the sediments are formed. The processes of weathering can be subdivided into chemical weathering, mechanical weathering, and organic weathering.

The three agents in the atmosphere which are responsible for chemical weathering are oxygen, carbon dioxide and water vapour. In certain cases, carbonic acid and excess water act on granite, albite, biotite, etc and give free silica, carbonate of alkali elements and other secondary minerals. This process loosens the rocks of land surface and alters them to easily erodible

material. Sometimes the iron ores are oxidized. In certain other cases hydration of minerals gives rise to new secondary minerals and increases their volume.

Several mechanical agents also disintegrate the parent rocks into fragmental material. These agents include:

- ♣ Freezing water
- ♣ Expansion caused by chemical changes
- ♣ Exfoliation resulting from sudden changes in temperature

Volume of water increases by approximately ten per cent when it freezes. When the water is confined, such expansion causes a considerable force. Therefore, water which enters the cracks and fissures and freezes tends to push the rock apart. Repeated freezing and thawing at high altitude can be one of the most important disintegrating forces for the rocks. In certain cases secondary minerals such as kaolinite formed by chemical weathering occupy a volume much greater than the original mineral. This can create forces which ultimately break the rocks. In some other processes called chemical exfoliation, the sheets are notably decayed and discolored. The work performed by the solutions which penetrate slowly along the cleavage cracks of crystals and between the grains and thereby induce the formation of new minerals of large volumes which causes disintegration. Exfoliation or scaling also occurs due to extreme heating followed by sudden cooling.

Organic agents of weathering are primarily burrowing animals and also roots and trunks of trees which wedge the rocks apart.

After the parent rocks are disintegrated, the material is transported from one place to another and deposited by streams, wind, or glaciers. The materials are called alluvium if transported and deposited by streams, loess if transported and deposited by wind, and glacial drift if it is transported and deposited by glaciers.

Stream erosion and deposition

The sediment load carried by streams comes from various sources. In the hilly areas the streams pick up coarse material from the talus. (Talus is a geological term describing a heap or sheet of coarse rock that has accumulated at the foot of the hill or on a steep slope.) Landslides also contribute to the load carried by streams. However, the major portion of the sediment load carried by streams comes from the erosion of material in the drainage basin; a certain amount also originates as a result of weathering of rocks from the bed and banks of the stream. The size of the sediment transported in any reach is dependent on the geology of the basin as well as the distance of the reach from the source. The amount of the sediment load carried depends on the size of materials, discharge, slope, and channel and catchment characteristics.

Wind erosion and deposition

Arid and semi-arid regions are characterized by relatively low and infrequent rainfall. As such, stream flows in these regions are small, unless, of course, the area is small and is traversed by stream flowing into the area from an area of appreciable rainfall. However, in general, stream erosion in arid and semi-arid regions is relatively small.

On the other hand, wind erosion becomes a predominant factor. High velocity winds, carrying fine sand with them, are effective agents of wind erosion. When wind blows over deserts and ploughed fields, fine sand and dust particles are carried away while the coarser material is left behind. This process is called deflation. The dust that is carried by the wind is transported to great distances. When the wind velocity is reduced, this material is deposited as loess.

Glacial erosion and deposition

As big pieces of ice move over land surface, they pick up loose rock, boulders, and sand along with them. This material acts as an abrasive agent to loosen other materials and reduce the size of this material. The transporting capacity of glaciers is large enough to transport rocks of the size of rooms. When the glacier melts most of the materials that it has been carrying is dropped and deposited.

2.2.3 Properties of sediment

Sediments are broadly classified as cohesive and noncohesive (or cohesionless). With cohesive sediments, the resistance to erosion depends on the strength of the cohesive bond binding the particles. Cohesion may far outweigh the influence of the physical characteristics of the individual particles. On the other hand, the cohesionless sediments generally consist of larger discrete particles than the cohesive soils. Cohesionless sediment particles react to fluid forces and their movement is affected by the physical properties of the particles such as size, shape, and density.

Sediment properties of individual particles that are important in the study of sediment transport are *particle size, shape, density, specific weight, and fall velocity*.

Size

Size is the basic and most readily measurable property of sediment. Size has been found to sufficiently describe the physical property of a sediment particle for many practical purposes. The size of a particle can be determined by *caliper, sieve or by sedimentation methods*. The size of an individual particle is not of primary importance in river mechanics or sedimentation studies, but the size distribution of the sediment that forms the bed and banks of the stream or reservoir are of great importance.

Shape

Shape refers to the geometric form or configuration of a particle regardless of its size or composition. Corey investigated several shape factors, and concluded that from the viewpoint of simplicity and effective correlation, the following ratio was most significant expression of shape.

$$S_p = \frac{c}{\sqrt{ab}} \quad (2.1)$$

In this equation, a, b, and c are the lengths of the *longest, the intermediate, and the shortest* mutually perpendicular axes through the particle, respectively, and S_p is the shape factor (also called Corey's shape factor). The shape factor is 1.0 for sphere. Naturally worn quartz particles have an average shape factor of 0.7.

Density

The density of a sediment particle refers to its mineral composition. Usually, specific gravity, which is defined as the ratio of specific weight or density of sediment to specific weight or density of water, is used as an indicator of density. Waterborne sediment particles are primarily composed of quartz with specific gravity of 2.65.

Specific weight

Specific weight is an important factor extensively used in hydraulics and sediment transport. The specific weight of deposited sediment depends on the extent of consolidation of the sediment. It increases with time after initial deposition. It also depends on the composition of the sediment mixture.

Fall velocity, ω

The fall velocity, or the terminal fall velocity, that a particle attains in a quiescent column of water, is directly related to relative flow conditions between the sediment particle and the water during conditions of sediment entrainment, transportation, and deposition. The fall velocity reflects the integrated result of size, shape, surface roughness, specific gravity, and viscosity of fluid. The fall velocity of a particle can be calculated from a balance between the particle buoyant weight and the resisting force resulting from fluid drag. The general drag equation is

$$F_D = C_D \rho A \frac{\omega^2}{2} \quad (2.2)$$

where F_D = drag force; C_D = drag coefficient; ρ = density of water; A = the projected area of particle in the direction of fall, and ω = the fall velocity.

The buoyant or submerged weight of a spherical sediment particle is

$$W_s = \frac{4}{3} \pi r^3 (\rho_s - \rho)g \quad (2.3)$$

Where, r is the particle radius.

The fall velocity can be solved from equations (2.2) and (2.3) once the drag coefficient has been determined. The drag coefficient is a function of Reynolds number and shape factor.

Theoretical consideration of drag coefficient: For a very slow and steady moving sphere in an infinite liquid at a very small Reynolds number, the drag force can be expressed as

$$F_D = 6 \mu \pi r \omega \quad (2.4)$$

The drag coefficient is then found to be (this is the viscous or Stokes range where R_e is less than 0.1)

$$C_D = \frac{24}{R_e} \quad (2.5)$$

Where, R_e = Reynolds number.

Equation (2.5) is acceptable for Reynolds numbers less than 1.0.

From equation (2.2) and (2.5), Stoke's equation can be obtained, i.e.,

$$F_D = 3 \pi d \rho \nu \omega \quad (2.6)$$

From equation (2.3) and (2.6), the terminal fall velocity for a sediment particle is

$$\omega = \frac{1}{18} \frac{\gamma_s - \gamma}{\gamma} g \frac{d^2}{\nu} \quad (2.7)$$

where d is the sediment diameter.

Equation (2.7) is applicable for the estimation of fall velocity of a sediment particle in water if the particle diameter is equal to or less than 0.1 mm. The value of kinematic viscosity in equation (2.7) is a function of water temperature, and can be computed from

$$\nu = 1.79^2 \times 10^{-6} / (1.0 + 0.033T + 0.00022T^2) \quad (2.8)$$

where T is water temperature in °C.

Oseen included some inertia terms in his solution of the Navier-Stokes equation. The solution thus obtained is

$$C_D = \frac{24}{R_e} \left(1 + \frac{3}{16} R_e \right) \quad (2.9)$$

Goldstein provided a more complete solution of the Oseen approximation, and the drag coefficient becomes

$$C_D = \frac{24}{R_e} \left(1 + \frac{3}{16} R_e - \frac{19}{1280} R_e^2 + \frac{71}{20480} R_e^3 + \dots \right) \quad (2.10)$$

Equation (2.10) is valid for Reynolds number up to 2.0.

Rubey's formula: Rubey introduced a formula for the computation of fall velocity of gravel, sand, and silt particles. For quartz particles with diameter greater than 1 mm, the fall velocity can be computed by

$$\omega = F \left[d g \left(\frac{\gamma_s - \gamma}{\gamma} \right) \right]^{1/2} \quad (2.11)$$

where the parameter F= 0.79 for particles greater than 1 mm settling in water with temperature between 10°C and 25°C, and d is the particle diameter.

For smaller grain sizes

$$F = \left[\frac{2}{3} + \frac{36 \nu^2}{g d^3 ((\gamma_s/\gamma) - 1)} \right]^{1/2} - \left[\frac{36 \nu^2}{g d^3 ((\gamma_s/\gamma) - 1)} \right]^{1/2} \quad (2.12)$$

For particle sizes greater than 2 mm, the fall velocity in 16°C water can be approximated by

$$\omega = 3.32 d^{1/2} \quad (\omega \text{ in m/s, } d \text{ in m}) \quad (2.13)$$

2.2.4 Bulk Properties of Sediment

The *size distribution, specific weight, and porosity* of bed material are the three important bulk properties in the study of sediment transport.

1. Particle Size Distribution: While the properties and behavior of individual sediment particles are of fundamental concern, the greatest interest is in groups of sediment particles.

Various sediment particles moving at any time may have different sizes, shapes, specific gravities, and fall velocities. The characteristic properties of the sediment are determined by taking a number of samples and making a statistical analysis of the samples to determine the mean, distribution, and standard deviation of the sample.

The most commonly used method to determine size frequency is mechanical or sieve analysis. In general, the results are presented as cumulative – size frequency curves. The fraction or percentage by weight of sediment that is smaller or larger than a given size is plotted against particle size.

Usually, sediments are referred to as *gravel, sand, silt or clay*. These terms refer to the size of the sediment particle. Table 2.1 presents the grain size scale of the American Geophysical Union. This scale is based on powers of 2, which yields a linear logarithmic scale via the phi-parameter defined as $\Phi = -\log_2 d$ (with d in mm).

Table 2.1 Grain size scale of American Geophysical Union

Class Name	Millimeters	Micrometers	Phi Values
Boulders	> 256		< -8
Cobbles	256 - 64		-8 to -6
Gravel	64 - 2		-6 to -1
Very coarse sand	2.0 - 1.0	2000 - 1000	-1 to 0
Coarse sand	1.0 - 0.50	1000 - 500	0 to +1
Medium sand	0.50 - 0.25	500 - 250	+1 to +2
Fine sand	0.25 - 0.125	250 - 125	+2 to +3
Very fine sand	0.125 - 0.062	125 - 62	+3 to +4
Coarse silt	0.062 - 0.031	62 - 31	+4 to +5
Medium silt	0.031 - 0.016	31 - 16	+5 to +6
Fine silt	0.016 - 0.008	16 - 8	+6 to +7
Very fine silt	0.008 - 0.004	8 - 4	+7 to +8
Coarse clay	0.004 - 0.0020	4 - 2	+8 to +9
Medium clay	0.0020 - 0.0010	2 - 1	+9 to +10
Fine clay	0.0010 - 0.00050	1 - 0.5	+10 to +11
Very fine clay	0.0005 - 0.00024	0.5 - 0.25	+11 to +12
Colloids	< 0.0024	< 0.24	> +12

Various methods are available to determine the particle size. Cobbles can be measured directly with ruler. Gravel, sand and silt are analyzed by wet or dry sieving methods yielding sieve diameters. Clay materials are analyzed hydraulically by using settling methods yielding the particle fall velocity from which the standard fall diameter is computed.

A natural sample of sediment particles contains particles of a range of sizes. The size distribution of such a sample is the distribution of sediment material by percentages of weight, usually presented as a cumulative frequency distribution (see Figure 2.1).

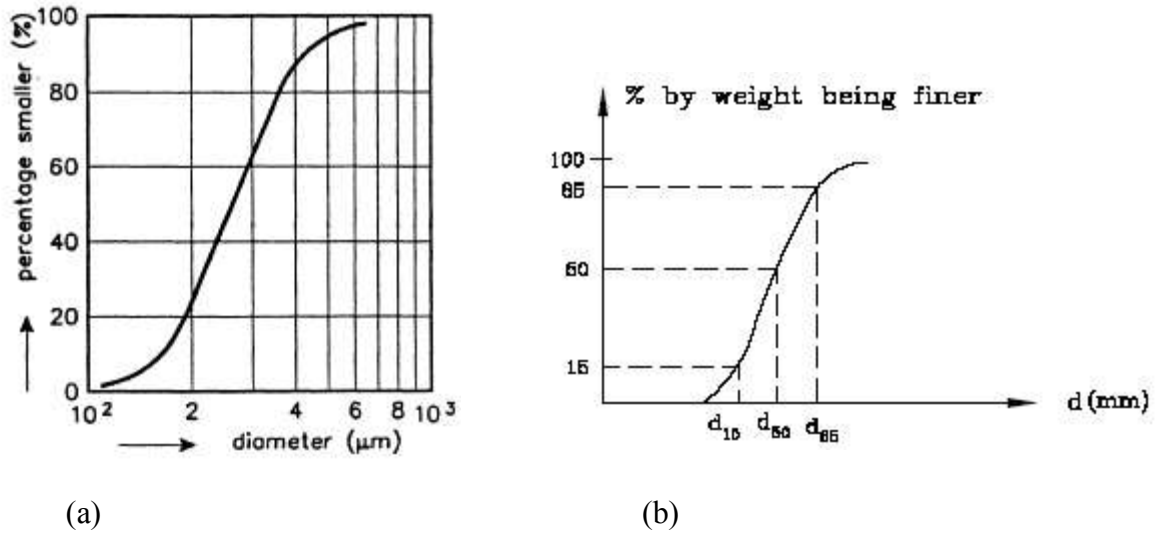


Figure 2.1 Particle size distribution curves

The frequency distribution is characterized by:

- ♦ **Median particle size** – d_{50} which is the size at 50% by weight is finer,
 - ♦ **Mean particle size** - $d_m = \sum (p_i d_i) / 100$, with p_i = percentage by weight of each grain size fraction, d_i .
 - ♦ **Geometric mean size** – $d_g = (d_{15.9} d_{84.1})^{1/2}$ is the geometric mean of the two sizes corresponding to 84.1% and 15.9% finer, respectively.
 - ♦ **Geometric standard deviation** – $\sigma_g = (d_{84.1} / d_{15.9})^{1/2}$.
 - ♦ **Gradation coefficient** – $G = \left(\frac{d_{84.1}}{d_{50}} + \frac{d_{50}}{d_{15.9}} \right)$
2. **Specific Weight:** The specific weight of deposited sediment depends on the extent of consolidation of the sediment. It increases with time after initial deposition. It also depends on the composition of the sediment mixture. The consolidation of sediment is of interest to hydraulic engineers. For example, the life of reservoirs varies as function of deposition and consolidation. Consolidation concepts can be used to convert sediment load, determined in units of weight, to volume of deposits in rivers, irrigation channels, etc.
 3. **Porosity:** Porosity is important in the determination of the volume of sediment deposit. It is also important in the conversion from sediment volume to sediment discharge. The following equation can be used for the computation of sediment discharge by volume including that due to voids, once the porosity and sediment discharge by weight have been determined, i.e.,

$$V_t = \frac{V_s}{1-p} \tag{2.14}$$

where V_t = total volume of sediment, including that due to voids, and V_s = volume of sediment excluding that due to voids.

Sediment Terminology

The science of sediment transport deals with the interrelationship between flowing water and sediment particles. An understanding of the physical properties of water and sediment particles is essential to our study of sediment transport. In this section, the most commonly used terminologies are introduced; fundamental properties of water and sediment particles are also given.

Some commonly used terms for describing the properties of water and sediment are:

- 3 **Density:** the mass per unit volume [kg/m^3]. The density of water is denoted by ρ while that of sediment is denoted by ρ_s .
- 4 **Specific weight:** the weight per unit volume [kN/m^3]. It is denoted by γ for water γ_s for sediment. The relationship between density and specific weight is
$$\gamma_s = \rho_s g \quad \text{for sediment, and}$$
$$\gamma = \rho g \quad \text{for water}$$
- 5 **Specific gravity:** The specific gravity, s , is the ratio of the specific weight of a solid or a liquid (a given material) to that of water at 4°C . The specific gravity of most common sediments is 2.65.
- 6 **Nominal diameter:** is the diameter of a sphere having the same volume as the particle.
- 7 **Sieve diameter:** is the diameter of a sphere equal to the length of the side of a square sieve opening through which the particle can just pass. As an approximation, the sieve diameter is equal to the nominal diameter.
- 8 **Fall diameter:** is the diameter of a sphere that has a specific gravity of 2.65 and has the same terminal fall velocity as the particle when each is allowed to settle alone in quiescent, distilled water. **The standard fall diameter** is the fall diameter determined at a water temperature of 24°C .
- 9 **Fall velocity:** is the average terminal settling velocity of a particle falling alone in quiescent distilled water of infinite extent. When the fall velocity is measured at 24°C , it is called the **standard fall velocity**.
- 10 **Angle of repose:** is the angle of slope formed by a given material under the conditions of incipient sliding.
- 11 **Porosity:** is a measure of the volume of voids per unit volume of sediment, i.e. $p = v_v/v_t$, where p = porosity, v_v = volume of voids, v_t = total volume of sediment, including that due to voids.
- 12 **Viscosity:** is the degree to which a fluid resists flow under an applied force. Dynamic viscosity is the constant of proportionality relating the shear stress and velocity gradient, i.e. $\tau = \mu du/dy$, where τ is shear stress, μ is dynamic viscosity and du/dy is the velocity gradient. **Kinematic viscosity** is the ratio between dynamic viscosity and fluid density, i.e. $\nu = \mu/\rho$, ν is kinematic viscosity.

2.3 Limitations of particle motion

The amount and size of sediment moving through a river channel are determined by three fundamental controls: *competence*, *capacity* and *sediment supply*.

Competence refers to the largest size (diameter) of sediment particle or grain that the flow is capable of moving; it is a hydraulic limitation. If a river is sluggish and moving very slowly it

simply may not have the power to mobilize and transport sediment of a given size even though such sediment is available to transport. So a river may be competent or incompetent with respect to a given grain size. If it is incompetent it will not transport sediment of the given size. If it is competent it *may* transport sediment of that size *if* such sediment is available (that is, the river is not supply-limited).

Capacity refers to the maximum amount of sediment of a given size that a stream can transport in traction as *bedload*. Given a supply of sediment, capacity depends on channel gradient, discharge and the calibre of the load (the presence of fines may increase fluid density and increase capacity; the presence of large particles may obstruct the flow and reduce capacity). *Capacity transport* is the competence-limited sediment transport (mass per unit time) predicted by all sediment-transport equations, examples of which we will examine below. Capacity transport only occurs when sediment supply is abundant (non-limiting).

Sediment supply refers to the amount and size of sediment available for sediment transport. Capacity transport for a given grain size is only achieved if the supply of that calibre of sediment is not limiting (that is, the maximum amount of sediment a stream is capable of transporting is actually available). Because of these two different potential constraints (hydraulics and sediment supply) distinction is often made between *supply-limited* and *capacity-limited transport*. Most rivers probably function in a sediment-supply limited condition although we often assume that this is not the case.

Much of the material supplied to a stream is so fine (silt and clay) that, provided it can be carried in suspension, almost any flow will transport it. Although there must be an upper limit to the capacity of the stream to transport such fines, it is probably never reached in natural channels and the amount moved is limited by supply. In contrast, transport of coarser material (say, coarser than fine sand) is largely capacity limited.

2.4 Incipient Motion

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. Cohesive forces are important when the bed consist of appreciable amounts of clay and silt particles.

The driving forces are strongly related to the local near-bed velocities. In turbulent flow conditions the velocities are fluctuating in space and time, which make together with the randomness of both particle size, shape and position that initiation of motion is not merely a deterministic phenomenon but a stochastic process as well.

Incipient motion is important in the study of sediment transport, channel degradation, and stable channel design. Due to the stochastic nature of sediment movement along an alluvial bed, it is difficult to define precisely at what flow condition a sediment particle will begin to move.

Let us consider the steady flow over the bed composed of cohesionless grains. The forces acting on the grain is shown in Fig.2.2.

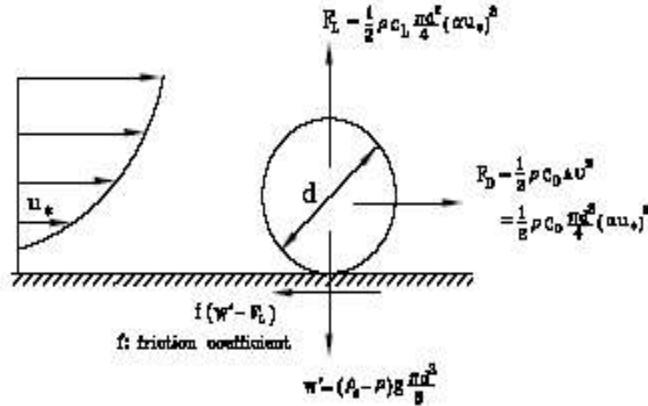


Figure 2.2 Forces acting on a grain resting on the bed.

The driving force is the flow drag force on the grain,

$$F_D = \frac{1}{2} \rho C_D \frac{\pi d^2}{4} (\alpha u_*)^2$$

Where the friction velocity u_* is the flow velocity close to the bed. α is a coefficient, used to modify u_* so that αu_* forms the characteristic flow velocity past the grain. The stabilizing force can be modeled as the friction force acting on the grain.

If $u_{*,c}$, critical friction velocity, denotes the situation where the grain is about to move, then the drag force is equal to the friction force. i.e.

$$\frac{1}{2} \rho C_D \frac{\pi d^2}{4} (\alpha u_{*,c})^2 = f \left((\rho_s - \rho) g \frac{\pi d^3}{6} - \frac{1}{2} \rho C_L \frac{\pi d^2}{4} (\alpha u_{*,c})^2 \right)$$

which can be re-arranged into

$$\frac{u_{*,c}^2}{(s-1) g d} = \frac{f}{\alpha^2 C_D + f \alpha^2 C_L} \frac{4}{3 \alpha^2}$$

Shields parameter is then defined as

$$\tau_* = \frac{u_*^2}{(s-1)g d} \tag{2.15}$$

We say that sediment starts to move if

$$u_* > u_{*,c} \quad \text{critical friction velocity } u_{*,c}$$

or $\tau_b > \tau_{b,c}$ critical bottom shear stress $\tau_{b,c} = \rho u_{b,c}^2$

or $\tau_* > \tau_{*,c}$ critical Shields parameter $\tau_{*,c} = \frac{u_{*,c}^2}{(s-1)g d}$

Fig.2.3 shows Shields experimental results, which relate $\tau_{*,c}$ to the grain Reynolds number defined as

$$Re = \frac{u_* d_n}{\nu} \tag{2.16}$$

The figure has 3 distinct zones corresponding to 3 flow situations

- 1) Hydraulically smooth flow for $Re = \frac{u_* d_n}{\nu} \leq 2$.
 d_n is much smaller than the thickness of viscous sublayer. Grains are embedded in the viscous sublayer and hence, $\tau_{*,c}$ is independent of the grain diameter. By experiments it is found $\tau_{*,c} = 0.1/Re$.

- 2) Hydraulically rough flow for $Re \geq 500$.
 The viscous sublayer does not exist and hence, $\tau_{*,c}$ is independent of the fluid viscosity. $\tau_{*,c}$ has a constant value of 0.06.

- 3) Hydraulically transitional flow for $2 \leq Re \leq 500$.
 Grain size is the same order as the thickness of the viscous sublayer. There is a minimum value of $\tau_{*,c}$ of 0.032 corresponding to $Re = 10$.

Note that the flow classification is similar to that of the Nikurase pipe flow where the bed roughness k_s is applied instead of d_n .

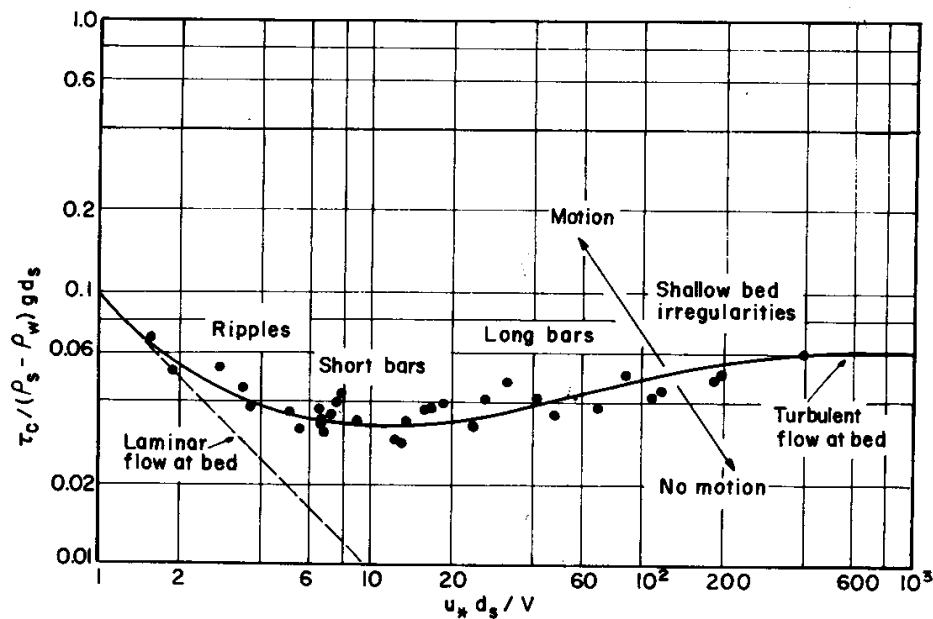


Figure 2.3 The Shields diagram giving $\tau_{*,c}$ as a function of Re (uniform and cohesionless grain).

2.5 Transportation mechanisms

The sediment load of a river is transported in various ways although these distinctions are to some extent arbitrary and not always very practical in the sense that not all of the components can be separated in practice:

1. Dissolved load
2. Suspended load

3. Intermittent suspension (saltation) load
4. Wash load
5. Bed load

Dissolved Load

Dissolved load is material that has gone into solution and is part of the fluid moving through the channel. Since it is dissolved, it does not depend on forces in the flow to keep it in the water column.

The amount of material in solution depends on supply of a solute and the saturation point for the fluid. For example, in limestone areas, calcium carbonate may be at saturation level in river water and the dissolved load may be close to the total sediment load of the river. In contrast, rivers draining insoluble rocks, such as in granitic terrains, may be well below saturation levels for most elements and dissolved load may be relatively small. Obviously, the dissolved load is also very sensitive to water temperature and other things being equal, tropical rivers carry larger dissolved loads than those in temperate environments. Dissolved loads for some of the world's major rivers are listed in table 2.2.

Dissolved load is not important to the geomorphologist concerned with channel processes because it is simply part of the fluid. On the other hand, dissolved load is very important to geomorphologists concerned with sediment budgets at a basin scale and with regional denudation rates. In many regions most of the sediment is removed from a basin in solution as dissolved load and must be accounted for in estimating erosion rates.

River	Drainage area, 10^3 km^2	Mean Discharge $10^3 \text{ m}^3 \text{ s}^{-1}$	Mean suspended Sediment load 10^6 t y^{-1}	Mean dissolved load 10^6 t y^{-1}	Percentage of total load carried in solution
AFRICA					
Congo	3500	41.1	48	37	44
Niger	1200	4.9	25	14	36
Nile	3000	1.2	2	12	86
Orange	1000	0.4	0.7	1.6	70
Zambezi	540	2.4	20	25	56
ASIA					
Ganges/Brahmaputra	1480	30.8	1670	151	8
Huang Ho	752	1.1	900	22	2
Indus	1170	7.5	100	79	44
Irrawaddy	430	13.6	265	91	26
Lena	2440	16.0	12	56	82
Mekong	795	21.1	160	59	27
Ob	2990	13.7	13	46	78
Yenisei	2500	17.6	15	60	80
AUSTRALIA					
Murray-Darling	1060	0.7	30	8.2	21
EUROPE					
Danube	817	6.5	83	53	39
Dnieper	527	1.7	2.1	11	84
Rhone	99	1.9	40	56	58
Volga	1459	7.7	27	54	67
NORTH AMERICA					
Columbia	670	5.8	14	21	60
Mackenzie	1810	7.9	100	44	31
Mississippi	3267	18.4	210	142	40
St Lawrence	1150	13.1	5.1	70	93
Yukon	840	6.7	60	34	36
SOUTH AMERICA					
Amazon	6150	200	900	290	24
Magdalena	260	6.8	220	20	8
Orinoco	990	36	150	31	17
Parana	2800	15	80	38	32

Table 2.2 Sediment loads of major world rivers (after Knighton, 1998: Figure 3.2). Sources: Degens et al (1991), Meybeck (1976) and Milliman and Meade (1983).

Suspended-sediment load

Suspended-sediment load is the clastic (particulate) material that moves through the channel in the water column. These materials, mainly silt and sand, are kept in suspension by the upward flux of turbulence generated at the bed of the channel. The upward currents must equal or exceed the particle fall-velocity (Figure 2.4) for suspended-sediment load to be sustained.

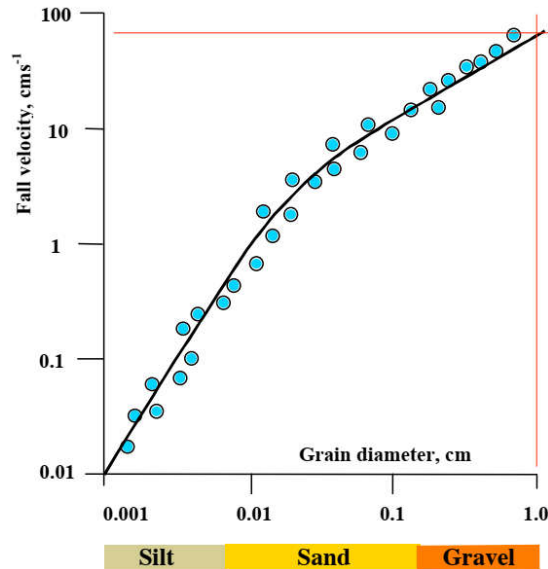


Figure 2.4 Fall velocity in relation to diameter of a spherical grain of quartz

Wash load

Although *wash load* is part of the suspended-sediment load it is useful here to make a distinction. Unlike most suspended-sediment load, wash load does not rely on the force of mechanical turbulence generated by flowing water to keep it in suspension. It is so fine (in the clay range) that it is kept in suspension by thermal molecular agitation (sometimes known as *Brownian motion*, named for the early 19th-century botanist who described the random motion of microscopic pollen spores and dust). Because these clays are always in suspension, wash load is that component of the particulate or clastic load that is “washed” through the river system.

Unlike coarser suspended-sediment, wash load tends to be uniformly distributed throughout the water column. That is, unlike the coarser load, it does not vary with height above the bed.

“Glacial flour” is a common source of wash load present in glacial rivers and lakes. In lakes deposition of the wash load may eventually occur but only after months of settling in an undisturbed pond or lake. The repeated annual deposit of wash load from glacial meltwater streams to lakes produces the “varved clays” that form the bottoms of glacial lakes. Because *all* clastic sediment in the water column is measured simultaneously with a suspended-sediment sampler, in practice it is not possible to distinguish between wash load and the rest of the suspended load.

Intermittently-suspended or saltation load

Some river scientists make the not especially useful distinction between fully-*suspended load* and *bed load* by classifying the intermediate and transient transport state as *saltation load* transport. These are particles that bounce along the channel, partly supported by the turbulence in the flow and partly by the bed. They follow a distinctively asymmetric trajectory (see Figure below).

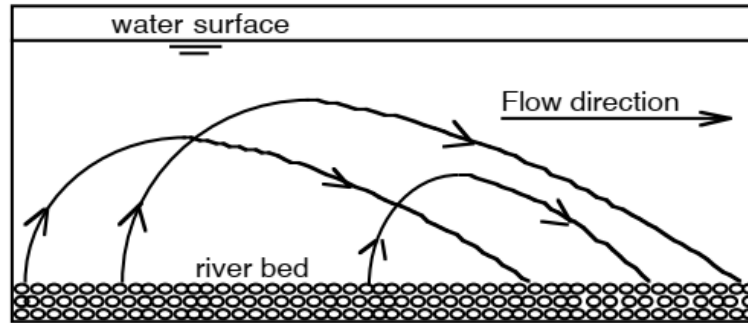


Figure 2.5 The trajectory of saltating (intermittently suspended) sediment grains moving in the flow

Saltation load may be measured as suspended load (when in the water column) or as bed load (when on the bed). Although the distinction between saltating load and other types of sediment load may be important to those studying the physics of grain movement, most geomorphologists are content to ignore it as a special case.

Bed Load (Traction Load)

Bed load is the clastic (particulate) material that moves through the channel *fully supported by the channel bed* itself. These materials, mainly sand and gravel, are kept in motion (rolling and sliding) by the shear stress acting at the boundary. Unlike the suspended load, the bed-load component is almost always capacity limited (that is, a function of hydraulics rather than supply). A distinction is often made between the *bed-material load* and the *bed load*.

Bed-material load is that part of the sediment load found in appreciable quantities in the bed (generally > 0.062 mm in diameter) and is collected in a bed-load sampler. That is, the bed material is the source of this load component and it includes particles that slide and roll along the bed (in bed-load transport) *but also those near the bed transported in saltation or suspension*. The material eroded from the river bed, referred to as *bed-material load*, is transported by the river as *bed load* and as *suspended load*. Coarse sediment (cobbles, gravel and coarse sand) is normally transported by the flow as bed load, close to the bed surface. Fine material (medium and fine sand) is transported in suspension. It is kept in suspension by the water turbulence for a distance that depends on the fall velocity of the grain

Bed load, strictly defined, is just that component of the moving sediment that is supported by the bed (and not by the flow). That is, the term “bed load” refers to a mode of transport and not to a source.

2.6 Bed forms and Bed Roughness

Bedforms are relief features initiated by the fluid oscillations generated downstream of small local obstacles over a bottom consisting of movable (alluvial) sediment materials. Once sediment starts to move, various bedforms occur.

Classification of Bed forms

1. Bedforms in Sand-bed Rivers

Many types of bedforms can be observed in nature. When the bedform crest is perpendicular (transverse) to the main flow direction, the bedforms are called transverse bedforms, such as ripples, dunes and anti-dunes (see Fig. 3.4). Ripples have a length scale smaller than the water depth, whereas dunes have a length scale much larger than the water depth. Figure 3.4 shows the various bedforms observed in rivers.

Ripples and dunes travel downstream by erosion at the upstream face (stoss side) and deposition at the downstream face (lee side). Antidunes travel upstream by lee side erosion and stoss side deposition (see Figure 2.6).

Bedforms with their crest parallel to the flow are called longitudinal bedforms such as ribbons and ridges. In laboratory flumes the sequence of bedforms with increasing flow intensity is

Flat bed \Rightarrow Ripples \Rightarrow Dunes \Rightarrow High stage flat bed \Rightarrow Antidunes

Plane (flat) bed: is a plane bed surface without elevations or depressions larger than the largest grain of the bed material.

Ripples: Ripples are formed at relatively weak flow intensity and are linked with fine materials, with d_{50} less than 0.7 mm. The size of ripples is mainly controlled by grain size. By observations the typical height and length of ripples are

$$H_r \approx 100d_{50} \quad L_r \approx 1000d_{50}$$

At low flow intensity the ripples have a fairly regular form with an upstream slope 6° and downstream slope 32° . Ripple profiles are approximately triangular, with long gentle upstream slopes and short, steep downstream slopes.

Dunes: The shape of dunes is very similar to that of ripples, but it is much larger. The size of dunes is mainly controlled by flow depth. Dunes are linked with coarse grains, with d_{50} bigger than 0.6 mm. With the increase of flow intensity, dunes grow up, and the water depth at the crest of dunes becomes smaller. It means a fairly high velocity at the crest, dunes will be washed-out and the high stage flat (plane) bed is formed.

Transition: This bed configuration is generated by flow conditions intermediate between those producing dunes and plane bed. In many cases, part of the bed is covered with dunes while a plane bed covers the remainder.

Antidunes: These are also called standing waves. When Froude number exceeds unity antidunes occur. The wave height on the water surface is the same order as the antidune height. The surface wave is unstable and can grow and break in an upstream direction, which moves the antidunes upstream.

Bars: These are bed forms having lengths of the same order as the channel width, or greater, and heights comparable to the mean depth of the generating flow. These are *point bars, alternate bars, middle bars, and tributary bars*.

Chutes and Pools: These occur at relatively large slopes with high velocities and sediment concentrations.

2. Bed forms in Gravel-bed Rivers

The bed materials of gravel-bed rivers usually have a broad range of grain sizes from sand particles to large boulders. As a result, selective transport processes and armouring of the bed surface may occur locally. The most common regime is the lower transport regime; the transition regime (with plane bed) is a rare event. Very regular bed-features such as mega-ripples and dunes have been observed in laboratory flumes and small-scale channels. Various types of three-dimensional bar forms (referred to as alternate bars, crescent bars, transverse bars, deltaic bars) have been observed in depositional regions of rivers (see Fig. 2.8).

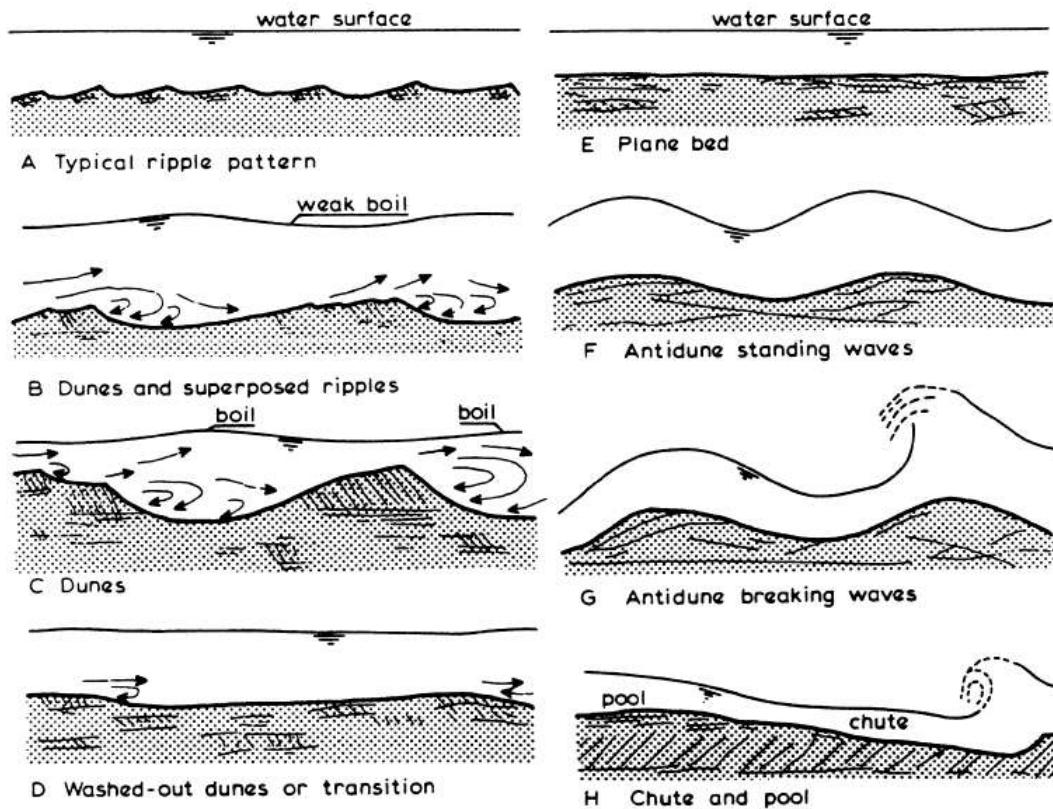


Figure 2.6 Bed form types in rivers

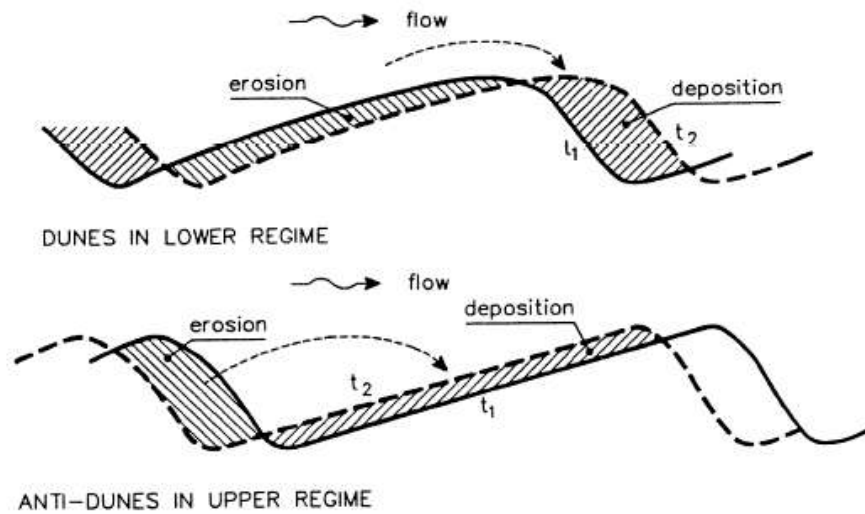


Figure 2.7 Bed form migration in lower and upper regimes

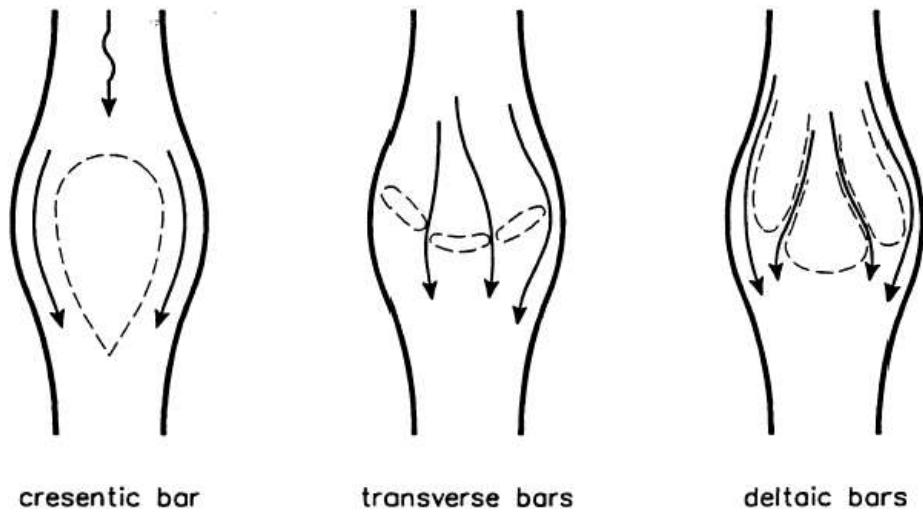


Figure 2.8 Three-dimensional large-scale bar forms

2.7 Bed Load, Suspended Load, Wash Load and Total Load Transport

General

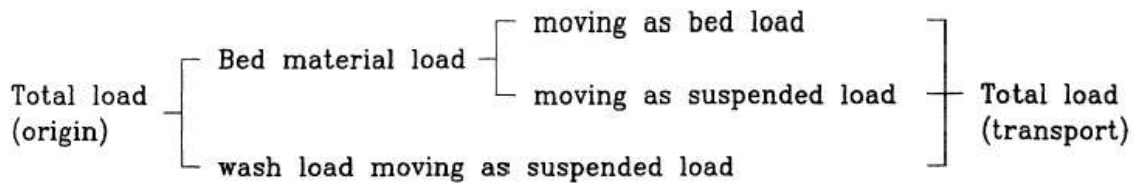
When the values of the bed shear velocity just exceeds the critical value for initiation of motion, the bed material particles will be rolling and/or sliding in continuous contact with the bed. For increasing values of the bed shear velocity the particles will be moving along the bed by more or less regular jumps, which are called saltations.

When the value of the bed shear velocity begins to exceed fall velocity of the particles, the sediment particles can be lifted to a level at which the upward turbulent forces will be of comparable or higher order than the submerged weight of the particles and as a result the particles may go into suspension.

Usually, the transport of particles by rolling, sliding and saltating is called *bed load transport*, while the suspended particles are transported as *suspended load transport*. The suspended load may also include the fine silt particles brought into suspension from the catchment area rather than from the streambed material (bed material load) and is called the *wash load*. A grain size of 63 μm (dividing line between silt and sand) is frequently used to separate between bed material and wash load.

Bed load and suspended load may occur simultaneously, but the transition zone between both modes of transport is not well defined.

The following classification and definitions are used for the total sediment transported in rivers.



- Bed material** : The material, the particle sizes of which are found in appreciable quantities in that part of the bed that is affected by transport.
- Bed material load** : The part of the total sediment transport which consists of the bed material and which rate of movement is governed by the transport capacity of the channel.
- Suspended load** : That part of the total sediment transport which is maintained in suspension by turbulence in the flowing water for considerable periods of time without contact with the stream bed. It moves with practically the same velocity as that of the flowing water.
- Bed load** : The sediment in almost continuous contact with the bed, carried forward by rolling, sliding or hopping.
- Wash load** : That part of the suspended load which is composed of particle sizes smaller than those found in appreciable quantities in the bed material. It is in near-permanent suspension and, therefore, is transported through the stream without deposition. The discharge of the wash load through a reach depends only on the rate with which these particles become available in the catchment area and not on the transport capacity of the flow.

2.7.1 Bed Load Transport

Usually, the transport of particles by rolling, sliding and saltating is called the bed load transport. Saltation refers to the transport of sediment particles in a series of irregular jumps and bounces along the bed (see Figure 2.9).

Bed load transport occurs when the bed shear stress, τ_0 exceeds a critical value $(\tau_0)_c$. In dimensionless terms, the condition for bed-load motion is:

$$\tau_* > (\tau_*)_c \quad \text{– Bed load transport}$$

where τ_* is the Shields parameter (i.e. $\tau_* = \tau_0 / (\rho(s-1)gd)$) and $(\tau_*)_c$ is the critical Shields parameter for initiation of bed load transport.

The sediment transport rate may be measured by weight (units: N/s), by mass (units: kg/s) or by volume (units: m³/s). In practice the sediment transport rate is often expressed per unit width and is measured either by mass or by volume.

Bed load, as one part of the bed material load, is often quantitatively small and hence does not represent a severe problem of sedimentation. On the other hand, as the main factor of the bed formation process, it is always of major importance. Roughness of alluvial channels is to a great extent determined by the movement of the bed load.

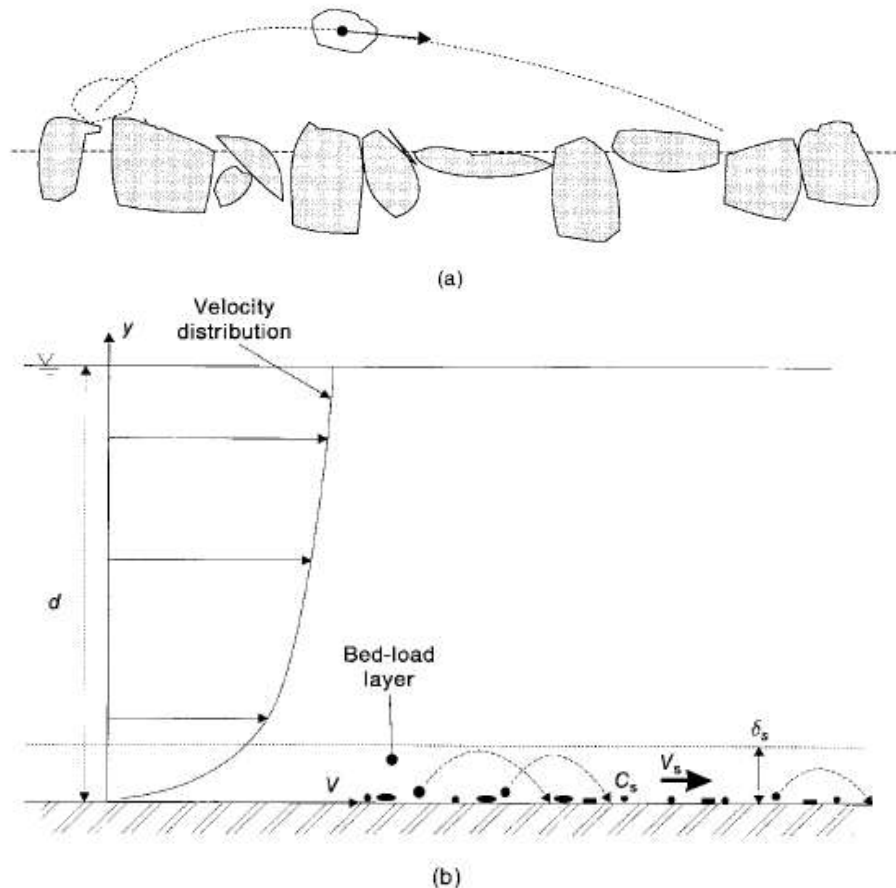


Figure 2.9 Bed load motion. (a) Sketch of saltation motion. (b) Definition sketch of the bed-load layer.

Bed Load Formulae

Various formulas are developed in the past for estimation of bed load discharge. Estimates of bed load transport using different formula for the same set of given data are also found to give widely different results. Here, only few of the most common formulae and approaches are presented.

1. Discharge Approach (bed load expressed in terms of discharge)

This approach, relating the quantity of transported sediment to water discharge, had been one of the main approaches adopted before the notion of shear stress gained prominence in later years. Main formulae of this kind are those proposed by Schoklitsch, Meyer-Peter and Casey, etc.

Schoklitsch Formula (1934 & 1943)

The stream is supposed to be wide, hence the sediment discharge relates to unit width of the bed. The main empirical relation reads as

$$q_b = 7000 \frac{S^{3/2}}{d^{1/2}} (q - q_c) \quad (2.17a)$$

where q_b = bed load (kg/s/m), d = grain size (mm), S = energy slope, q = specific water discharge ($m^3/s/m$), q_c = critical discharge ($m^3/s/m$).

The critical water discharge, i.e. the discharge that causes incipient motion, for sediments with specific gravity 2.65 is given by

$$q_c = 1.94 \times 10^{-5} \frac{d}{S^{3/5}} \quad (2.17b)$$

The 1943 Schoklitsch formula is

$$q_b = 2500 S^{3/2} (q - q_c) \quad (2.18a)$$

For sediments with specific gravity 2.65, the critical discharge in eq (2.18a) is given by

$$q_c = \frac{0.6 d^{3/2}}{S^{1/6}} \quad (2.18b)$$

Equations (2.17) and (2.18) have been developed for uniform grain distributions. It is, however, generally applied to non-uniform distributions also, taking d_{50} (median size) as the characteristic diameter for the mixture. Schoklitsch also suggested a more accurate method, which is as follows:

Sediment mixture is arbitrarily subdivided into several size sub-ranges having mean diameters d_a, d_b, d_c, etc ; and the partial quantity of each sub-range is then determined and expressed as percentage of the total quantity. Subsequently, partial bed loads $q_{ba}, q_{bb}, q_{bc}, etc$ are computed for each mean diameter for the given discharge, q , and given slope, S , using equations (2.17 and 2.18). The total bed load for the sediment mixture is then obtained,

$$q_b = a q_{ba} + b q_{bb} + c q_{bc} + \dots$$

where a, b, c, \dots indicate the percentage quantities that each partial sub-range is of the total.

This formula is applicable to grain sizes in the range of 0.3 – 7.0 mm.

2. Shear Stress Approach

This approach is much more favored today, because of the importance accorded to the shear stress in all aspects of the sediment movement in alluvial channels. Formulae of this type are those of Straub-Du Boys, Shield, Kalinske, Meyer-Peter and Mueller, etc.

The best known of these, and probably the most widely used, is the Meyer-Peter and Mueller formula; it also gives the best agreement with measured data.

The maximum amount of sediment that is transported by the river as bed-material load can be related to the stream velocity and is called *sediment transport capacity*. Therefore, the transport of material eroded from the riverbed is *capacity limited*, which means that a certain current cannot transport more sediment than a certain amount, even when this is available on the river bed.

In alluvial bed rivers, the bed-material load can be assumed to be always equal to the sediment transport capacity. This is not true, however, for rivers that are not fully-alluvial, i.e. those rivers characterized by rock outcrops or armoured bed. For semi-alluvial rivers the amount of bedmaterial load is generally smaller than the sediment transport capacity, because of the limited availability of sediment from the bed. All transport capacity formulas have been derived from laboratory experiments and include empirical features. A relevant consequence is that all morphodynamic predictions that are based on sediment transport formulas have an empirical character and are not exact. The general sediment transport capacity formula has the form:

$$q_s = m(u - u_c)^b \quad (2.19)$$

where q_s is the amount of bed-material load per meter of river width (m^2/s); m is a proportionality coefficient, u is the local stream velocity; u_c is the critical value of velocity below which no sediment transport occurs; b is the exponent (values always larger than 3).

The best formula is always the one that was derived for flow and sediment conditions that are similar to the case under study. In general, the presence of a threshold for sediment motion (for instance u_c in Equation above) is important for coarse sand, gravel and large sediment particles. It is less relevant for fine sand, which is almost always mobilized by river flows.

Meyer-Peter and Mueller Formula

A commonly used formula for **gravel-bed rivers** is the one derived by Meyer-Peter & Müller (1948). This formula (MPM) is restricted to bed load capacity, since the conditions for which it was derived did not present much transport in suspension::

$$q_s = K_{MPM} \sqrt{\Delta g D_m^3} \sqrt{(\mu\theta - 0.047)^3} \quad (2.20)$$

where

q_s	= amount of sediment transported as bed load per meter of river width (m^2/s)
D_m	= mean sediment grain size (m).
$\mu = \sqrt{(C/C')^3}$	= bed form factor (-)
C	= Chézy coefficient ($m^{1/2}/s$)
C'	= grain related Chézy coefficient: $C' = 18 \log_{10} \left(\frac{12h}{D_{90}} \right)$ with D_{90} grain size of 90% passing (m), also called C_{90}

The sediment transport formula computes the volume of sediment (bed-material load) transported per unit of time and per metre of river width, in m^2/s . If this volume includes an extra 40% occupied by pores (sand): $K_{MPM} = 13.3$. If the volume does not include the volume occupied by pores: $K_{MPM} = 8$. Wong and Parker (2008) recently corrected the value of K_{MPM} . According to them, $K_{MPM} = 6.65$ (with pores) and $K_{MPM} = 4$ (without pores). They also suggest imposing $\mu = 1$ if bedforms (dunes) are absent.

Unlike Engelund & Hansen, the Meyer-Peter & Müller formula includes a threshold. If $\theta\mu < 0.047$ no sediment transport occurs. Thus $\theta\mu = 0.047$ represents Meyer-Peter & Müller's condition for initiation of motion. This threshold is not coincident with the Shield's threshold, because it was derived from other experimental data. The MPM threshold is valid only for the experiments carried out by Meyer-Peter & Müller and for river channels satisfying the criteria for the application of the formula:

$$\frac{\omega_s}{u_*} > 1$$

$$D_m > 0.4mm$$

$$\pi\theta < 0.2$$

$$b = \frac{3\mu\theta}{\mu\theta - 0.047}$$

This equation shows that the exponent b varies with the stream velocity and is high close to the conditions of initiation of motion. Values of b approaching 3 reflect conditions of high sediment mobility in which sediment is partly moving in the water column.

Shields Formula

The semi-empirical formula derived by Shields for a level bed is

$$q_b = 10 \frac{q \gamma S (\tau - \tau_c)}{\gamma_s (\gamma_s - \gamma) d} \quad (2.21)$$

where $d = d_{50}$, and $S =$ bed slope.

In this formula τ and τ_c can be calculated from

$$\tau_c = 0.056(\rho_s - \rho)gd_{50} \quad \text{and}$$

$$\tau = \rho g R S$$

Equation (2.21) is dimensionally homogeneous, and can be used for any system of units. The critical shear stress can also be obtained from Shields diagram.

2.7.2 Suspended Load Transport

Suspended load refers to sediment that is supported by the upward components of turbulent currents and stays in suspension for an appreciable length of time. In most natural rivers, sediments are mainly transported as suspended load.

The suspended load transport can be defined mathematically as

$$q_{sv} = \int_a^h \bar{u} \bar{c} dz \quad (2.22a)$$

$$q_{sw} = \gamma_s \int_a^h \bar{u} \bar{c} dz \quad (2.22b)$$

where q_{sv} and q_{sw} are suspended load transport rates in terms of volume and weight, respectively; \bar{u} and \bar{c} are time averaged velocity and sediment concentration, by volume at a distance z above the bed, respectively; a is thickness of the bed load transport; and h is the water depth.

Before eq. (2.22) is integrated, \bar{u} and \bar{c} must be expressed mathematically as a function of z . Under steady equilibrium conditions, the downward movement of sediment due to the fall velocity must be balanced by the net upward movement of sediment due to turbulent fluctuations, i.e.,

$$\omega C + \varepsilon_s \frac{dC}{dz} = 0 \quad (2.23)$$

where ε_s is the momentum diffusion coefficient for sediment, which is a function of z ; ω is fall velocity of sediment particles; and C is sediment concentration.

For turbulent flow, the turbulent shear stress can be expressed as

$$\tau_z = \varepsilon_m \rho \frac{du}{dz} \quad (2.24)$$

where ε_m is kinematic eddy viscosity of fluid or momentum diffusion coefficient for fluid. It is generally assumed that

$$\varepsilon_s = \beta \varepsilon_m \quad (2.25)$$

where β is a factor of proportionality.

For fine sediments in suspension, it can be assumed that $\beta = 1$ without causing significant error. Eq.(2.23) can also be written as

$$\frac{dC}{C} + \frac{\omega dz}{\varepsilon_s} = 0 \quad (2.26)$$

and integration of eq.(3.43) yields

$$C = C_a \exp\left(-\omega \int_a^z \frac{dz}{\varepsilon_s}\right) \quad (2.27)$$

where C and C_a are sediment concentrations by weight at distance z and a above the bed, respectively.

The shear stress at a distance z above the bed is

$$\tau_z = \gamma S (h - z) = \tau \left(1 - \frac{z}{h}\right) \quad (2.28)$$

where τ and τ_z are shear stresses at channel bottom and a distance z above the bed, respectively.

Assume the Prandtl – von Karman velocity distribution is valid, i.e.,

$$\frac{du}{dz} = \frac{U_*}{k z} \quad (2.29)$$

From equations (2.24), (2.28) and (2.29),

$$\epsilon_m = k U_* \frac{z}{h} (h - z) \quad (2.30)$$

and $\epsilon_s = \beta k U_* \frac{z}{h} (h - z) \quad (2.31)$

Equation (2.30) indicates that $\epsilon_m = 0$ at $z = 0$ and $z = h$. The maximum value of ϵ_m occurs at $z = \frac{1}{2} h$.

On substituting eq. (2.31) into eq. (2.26) and integration from a to z , assuming $\beta = 1$, yields

$$\frac{C}{C_a} = \left(\frac{h - z}{z} \frac{a}{h - a} \right)^Z \quad (2.32)$$

where $Z = \frac{\omega}{k U_*}$ is known as the Rouse constant and equation (2.32) is called the Rouse equation. This equation gives the distribution of the suspended sediment concentration over the vertical for various values of Z (see Fig. 2.10).

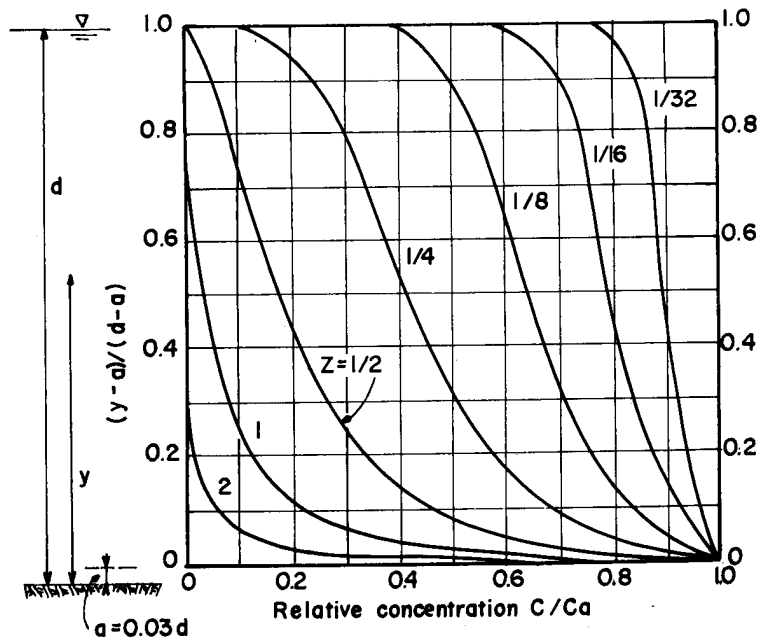


Figure 2.10 Suspended sediment distributions according to equation (2.32)

2.7.3 Wash load

Silt and clay that is transported by river flows mainly originates from soil erosion (due to rain, landslides, debris and mud flows) and from the falling of cohesive river banks. This material is transported as suspended load. It is kept in suspension by the water turbulence and since the fall velocity is very small, it does not settle in the river main channel, but flows downstream until it reaches a lake, a reservoir or the estuary. It is called *wash load*. The amount of material transported as wash load is *supply limited*, which means that the amount of this very fine material travelling with the water is generally limited by its input and not by the capacity of the current to transport it. Extreme high concentrations of fine sediment might transform the water flow into a mud flow. In general, there is no quantitative relation between flow velocity and wash load, except from some specific cases, although the highest amounts of wash load often occur during floods. This means that the amount of wash load cannot be determined as a function of the flow characteristics, such as flow velocity. It has to be determined by other means, i.e. from soil erosion rates or by measuring suspended solid concentrations for long periods of time.

High concentrations of sediment flowing on steep slopes (steepness higher than 15%) are called *debris flows*. This sediment is a mixture of large and small sediment grains. If a debris flow reaches the valley and enters the river bed, where the slope is much gentler, most of the largesized material settles and only a limited part is transported by the river current as bed load (capacity limited), while the very fine material continues its journey to the sea as wash load.

Organic sediment transport is fundamentally different from inorganic transport. Organic particles have low specific density, with relative density in water close to unit, and surface area-to-volume ratios that is higher than inorganic sediments. Besides, organic material can degrade into soluble products. Relationships between organic sediment concentration and discharge vary considerable among watersheds due to differences in their sources of organic materials. Large rivers with relatively long flood pulses receive important inputs of organic matter from their floodplains as they sweep back and forth across the aquatic-terrestrial transition zone during floods.

2.7.4 Total Load Transport

Based on the mode of transportation, total load is the sum of bed load and suspended load. Based on the source of material being transported, total load can also be defined as the sum of bed material load and wash load. Wash load consists of fine materials that are finer than those found in the bed. The amount of wash load depends mainly on the supply from the watershed, not on the hydraulics of the river. Consequently, it is difficult to predict the wash load based on the hydraulic characteristics of a river. Most total load equations are, therefore, totals bed material load equations.

General Approaches

There are two general approaches to the determination of total load:

- (1) Computation of bed load and suspended load separately and then adding them together to obtain total load – indirect method, and
- (2) Determination of total load function directly without dividing it into bed load and suspended load – direct method.

Engelund & Hansen (bed + suspended bed material load → total load)

This formula provides the **total load** thus, direct method. A commonly used sediment transport capacity formula is the one designed by Engelund & Hansen (1967), which is valid for **sand-bed rivers only**. According to Engelund & Hansen $b = 5$ and $u_c = 0$; m is given by the following expression:

$$m = \frac{1}{K_{EH} C^3 \Delta^2 D_{50} \sqrt{g}} \quad (2.33)$$

In the formula:

- C = Chezy's coefficient ($m^{1/2}/s$)
- Δ = relative density of sediment in water (-)
- ρ_s = sediment density (kg/m^3)
- ρ = water density (kg/m^3)
- D_{50} = median sediment grain size (50% of sediment material has smaller diameter) (m)
- g = acceleration due to gravity (m/s^2).

The sediment transport formula computes the volume of sediment (bed-material load) transported per unit of time and per metre of river width, in m^2/s . If the volume of sediment includes also an extra 40%, i.e. the volume occupied by the pores between the grains (sand): $K_{EH} = 12$. If the volume does not include pores: $K_{EH} = 20$. The sediment balance equation, which is necessary to compute the bed level changes in a specific reference area, will have to take into account whether the adopted sediment transport formula includes pores or not.

The sediment transport formula of Engelund & Hansen was derived from cases in which sediment (sand) was transported as bed and suspended load. Wash load is not taken into account. This formula can be applied to sediments and flows fulfilling the following criteria:

$$\frac{\omega_s}{u_*} < 1$$

$$0.19 < D_{50} < 0.93mm$$

$$0.07 < \theta < 6$$

in which:

- ω_s = fall velocity of sediment (m/s)
- $u_* = u \sqrt{\frac{g}{C^2}}$ = bed shear velocity (Eq. 7) (m/s)
- $\theta = \frac{u^2}{C^2 \Delta D_{50}}$ = Shields parameter (Eq. 8) (-).