MS698–3: Sediment transport processes in coastal environments
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Lecture 1: Introduction and Sediment properties

- Introduction to sediment transport processes.
- Physical properties of sediment grains: size, shape.
- Physical properties of sediment bed: size distribution, porosity.
- Physical properties of suspensions: size distribution, concentration.

Source materials

- Raudkivi, A.J.; **Loose Boundary Hydraulics**, Chapter 2: Sediment Properties. Third Edition; Pergamon Press, Oxford, 538pp.
- Rijn, Leo C. van; **Principles of sediment transport in rivers, estuaries, and coastal seas**, Section 3: Fluvial and sediment properties. Aqua Publications, 1993.

Handouts

- Sediment sizes. Figure 3.1 from van Rijn; Shape parameters. Folk, 1955
- Example size distribution from New York Bight.

Class business

- Class: 11:30 – 12:20 in Classroom C on Tuesdays and Thursdays.
- Office hours: 11:30 – 1:00 Mondays or by appointment.
- The reading for next Thursday (Jan. 23) can be downloaded from Elsevier: Drake, D.E., R. Eagenhouse, W. McArthur. 2002. Physical and chemical effects of grain aggregates on the Palos Verdes margin, southern California. *Continental Shelf Research* 22: 967–986.

Introduction to sediment transport processes.

Sediment is defined as fragmented material formed by physical and chemical weathering of rocks. Sediment transport study includes movement of huge boulders down mountain sides, to diffusion of colloid-sized material in groundwater systems. Transport is driven by gravity, and drag forces between the sediment and surrounding fluid (air or water). This course concentrates on sands, silts, and clays typically mobilized in coastal environments.

Transport of sediment is usually divided into three types: bedload, saltation, and suspension. Bedload transport is defined as transport where sediment grains roll or slide along the sediment bed. Saltation is defined as transport where individual grains hop along the bed, but reach heights of a few grain diameters above the bed so that they loose contact with bed material. During more intense flows, sediment grains reach sufficient height above the bed that turbulent mixing is of the same order of magnitude as gravitational settling. Sediment can then be suspended high above the bed and transport might be dominated by the suspended mode.

Why study sediment transport processes within coastal sciences?

- Earth surface is covered with sediment. The mechanics of transport determine surface morphology.
- Records of active transport are recorded in sedimentary rocks. An understanding of transport mechanisms is critical for interpreting such deposits.
- Many persistent contaminants become associated with sediment; understanding sediment transport is then necessary for predicting contaminant fate and transport pathways.
- Transport processes can impact on people (erosion, landslides).
- Many people live near the coasts and rely on seafood.
- Coastal regions are the interface between terrestrial and deep sea.
- Much of the stratigraphic record stems from nearshore, shelf, estuarine, and epicontinental seas.
- Challenges in coastal sediment transport include the fact that such environments are difficult to observe, and that significant transport occurs during infrequent, large-magnitude events.
- Advances in instrumentation continue to facilitate improved understanding of the physical processes that transport sediment.
- While much sediment transport research has been conducted using civil engineering and empirical approaches, efforts that are more physics, or process-based continue to make strides. Some advantages of a physics-based approach are that the assumptions made in a truly physical are stated explicitly. This provides a more firm base for prediction and for extension of the approach to diverse environments.

Physical properties of sediment grains.

Sediment transport depends on sediment properties, characteristics of the sediment bed, and properties of the fluid flow. In this lecture we'll cover the properties of individual sediment grains and sediment beds that impact transport.

The properties of individual sediment grains include sediment size (D or ϕ), sediment density (ρ_s), shape, and chemical composition.

Sediment density (ρ_s) depends on the mineralolgy of the sediment. Much natural sediment in coastal environments is made up of quartz, because quartz is abundant on the earth's crust and resists chemical and physical weathering. The density of quartz, $\rho_s = 2.65 g/cm^3 = 2650 kg/m^3$, is often assumed for sediment density in transport formulas. While many beaches are predominantly quartz and feldspar, they will often have some fraction of heavy minerals, including hornblende, garnet, and magnetite. These minerals have densities as high as $\rho_s = 10 g/cm^3$. Carbonate sands make up beach material off of some areas. Carbonates tend to be transported in a manner similar to quartz, but they will undergo chemical weathering. Other minerals such as montmorillonite, bentonite, and kaolinite are often found as clay-sized particles, and can make up a substantial part of the sediment bed in estuaries. Mud refers to mixtures of silts and clays. Sediment grain size can vary over several orders of magnitude in coastal environments. See the copy of AGU's size classifications for a more rigorous breakdown of these:

Type	Size Range (mm)	Size Range (ϕ)	
Gravel	>2 mm	$<$ -1 ϕ	Some beaches, fluvial
Sand	0.063 – 2 mm	4 – -1 ϕ	beaches, shelves, estuaries
Silt	0.004 – 0.063 mm	8 – 4 ϕ	outer shelves, estuaries
Clay	$<$ 0.004 mm	$>$ 8 ϕ	some shelves, estuaries

Those of us who study fine-grained sediments have to be able to convert between at least three types of units. Grain size usually refers to grain diameter; and is reported as millimeters (mm), micro-meters (microns or $\mu m = 10^{-6} m =$ 1 mm/1000), or "phi" units (ϕ). The phi size scale is logarithmically spaced, and grain diameter increases as phi size decreases. To convert between phi (ϕ) and diameter (D) in mm:

$$\begin{aligned} \phi &= -\log_2 \left(\frac{D}{1mm} \right) \\ \phi &= -\frac{\log\left(\frac{D}{1mm}\right)}{\log 2} \qquad (1) \\ 2^{-\phi} &= D(mm). \qquad (2) \end{aligned}$$

Different measurement techniques are used for different types of sediment; from manual methods for coarse-grained environments, to sieving for gravels and sands, to more intense laboratory methods for fine-grained material. Each sediment grain also has a range of dimensions that could be thought of as a "diameter", including the long (L), short (S), and intermediate (I) diameters. These give rise to biases in specification of grain size, as well as different definitions of grain diameter.

- The *sieve diameter* represents the diameter of the smallest circle that encompasses one dimension of the grain. It is the scale of the sieve mesh that would trap the sediment. It would be close to S and I.

- The *nominal diameter* represents the diameter of the sphere that would take up the same volume as the sediment grain. It would be somewhere in between L and S.

- The *standard fall diameter* represents the diameter of the quartz sphere that would settle at the same speed as the sediment grain in still, distilled water at 24^o/C. It arises from settling type measures of grain size.

A useful technique in fluid mechanics is to non-dimensionalize parameters. Grain diameter is often non-dimensionalized using an argument based on a force balance between the weight of the fluid and the drag that would be created around it. Non-dimensional grain diameter is usually specified as D_*, though van Rijn (equation 3), and J.D. Smith and Dietrich (equation 4) use different definitions; they are similar, and most applications specify which D_* they use.

$$D_* = \left[\frac{(\rho_s - \rho)\, g}{\rho \nu^2}\right]^{1/3} D \tag{3}$$

$$D_* = \left[\frac{(\rho_s - \rho)\, g}{\rho \nu^2}\right] D^3. \tag{4}$$

Here, g is acceleration due to gravity, and ν is the viscosity of the fluid.

Grain *shape* refers to the *sphericity* and *roundness* or *angularity* of a sediment grain. Shape and angularity impact sediment bed packing, the roughness of the sediment bed, settling rates of sediment, and bed friction. Sediment shape depends on the type of weathering processes that formed the sediment, and the transport history of the sediment. Sphericity refers to how equidimensional the grain axes are; while roundness or angularity refers to how sharp the angles of the grain are. The *Corey Shape Factor* (CSF) reports the sphericity of the grain as $CSF = S/\sqrt{IL}$; where L is the longest interior diameter of the grain, and S and I are the short and intermediate diameters that are perpendicular to L. "Naturally shaped" sediment tends to have CSFs of about 0.7. The roundness/angularity is usually defined by visual inspection and is defined on the *Powers scale* of P= 0 to 6; where very angular sediment has a low Powers scale and rounded sediment has a high value. "Naturally worn" sediments typically have Powers index of about 3.5 (I think; unverified).

Physical properties of sediment bed.

Sediment *texture* of a location is defined by the size distribution, density, shape, and sortedness of the sediment. Grain size distributions describe the population of sediment sizes found at a particular time and place. They can be plotted as histograms that describe the weight fraction (m_i) of sediment found within different ranges of grain diameter (D_i). When plotted on a log-scale, many natural grain size distributions display normal (i.e. Gaussian) distributions.

This is one reason that phi units have been used. The grain size distribution is summarized by an identification of a characteristic grain diameter, and often by the standard deviation around this characteristic. The "middle" of the size distribution can be described as the *mode* (size class of highest frequency), the *median* (D_{50}; half of the sediment is finer than this size), the *arithmetic mean* (weighted average of the grain size, $D_{mn} = \sum_i D_i m_i / \sum_i m_i$), or the *geometric mean* (weighted average of phi size $\phi_{mn} = \sum_i \phi_i m_i / \sum_i m_i$). The arithmetic mean will always be larger than the geometric mean; and if the sediment has a log-normal distribution, then the geometric mean will equal the median (D_{50}). A cumulative distribution function (C.D.F.) can also be used to identify the median grain size, and other important measures of grain size (D_{16}, D_{84} etc.) where x% of the sediment is finer than D_x.

The *standard deviation*, σ of a grain distribution indicates how well-sorted the grain distribution is, and this can provide clues as to the transport mechanisms that have operated on the sediment. An estimate of the standard deviation can be obtained from the cumulative distribution function; $\sigma \approx 0.5\,(D_{84} - D_{16})$. A $\sigma = 1\phi$ implies that D_{84} is twice as large as the median grain size. The sortedness of a grain distribution is defined as

$\sigma = 0 - 0.5\ \phi$	Well sorted
$\sigma = 0.5 - 1\ \phi$	Moderately sorted
$\sigma = 1 - 2\ \phi$	Poorly sorted
$\sigma > 2\ \phi$	Very poorly sorted

An important consideration in sediment transport is the relative volume occupied by sediment within a sediment bed. This is quantified using *porosity* (n) or bed concentration ($c_b = 1 - n$). The unitless bed concentration is the volume fraction of sediment within a unit volume of bed; and porosity is the volume of pore-space within the unit volume of bed. For sandy sediments (regardless of grain size), porosity seldom approaches the limit of well-packed spheres ($n = 0.26$), but does approach the value for randomly packed spheres ($n = 0.36 - 0.4$). Porosity does not depend on grain size, is sensitive to sediment sorting. Well sorted (non-cohesive) material will have porosities of about 0.45. Poorly sorted material attains lower porosities, because fine-sized grains can fill voids left by larger grains. Porosity of natural, non-cohesive, beds is often assumed to equal $n = 0.4$ for such sediments.

Porosity in muddy (cohesive) beds varies widely. Newly deposited muds can have very high porosity ($n \approx 0.8 - 0.95$), but porosity may decrease if the deposit dewaters as it ages.

Sediment transport formulas use many conventions for specifying density. Different schools of research use different symbols, so always make sure that the units work out when you try out a new formula.

- Density of the sediment grains is ρ_s. Quartz density is $\rho_s = 2.65 g/cm^3$.
- Density of the water is ρ; remember that this is somewhat sensitive to temperature and salinity. Typical seawater is $\rho \approx 1.025 g/cm^3$.

- The *bulk density* or *dry density* of a sediment is the dried mass per unit volume of a sediment bed, so it accounts for porosity and has units of mass/volume ($\rho_b = c_b\rho_s = (1-n)\,\rho_s$). Bulk density of a sandy bed might be about 1.7 $g/cm^3 = 1700\ kg/m^3$, and of a high-water content mud only 0.5 g/cm^3.

- Sometimes a non-dimensionalized density is used, defined as the *specific gravity* or *relative sediment density*, where sediment density is normalized by water density ($s = \rho_s/\rho$).

- Some schools of thought use the *submerged specific sediment density*, which accounts for the buoyancy of the particle and is non-dimensionalized by the fluid density; $R = \frac{(\rho_s-\rho)}{\rho}$.

- Other times, the formula needs the fluid-sediment mixture density; $\rho n + \rho_s c_b \;=\; \rho\,(1+(s-1)\,c_b)$.

Physical properties of suspensions.

Suspended sediments are described using similar characteristics as those listed for bed sediments (size distribution, mean size, etc.). The amount of sediment in suspension is quantified as the *concentration* to represent the amount of sediment per unit volume of fluid. Concentration can be reported as a volumetric concentration (v_s/v_{H_2O}, dimensionless) or mass concentration (m_s/v_{H_2O}, g/L, mg/L, etc.). Concentrations in coastal environments can range from near-zero to several hundred mg/L and higher. Fluid muds are often categorized as suspensions that exceed 10 g/L.