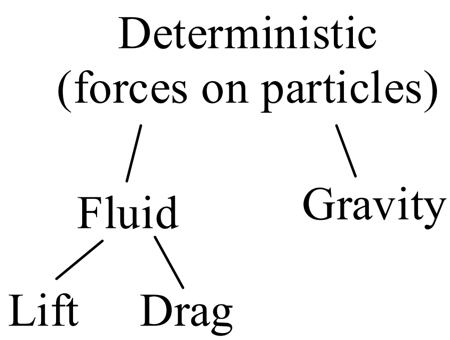
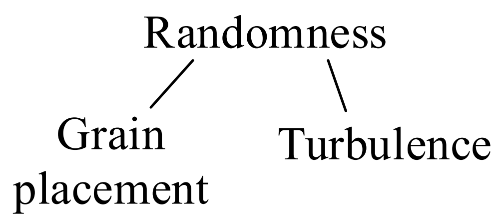
Initiation of Motion

The initiation of movement of individual grains is dependent on a variety of factors, both deterministic and random.





Many models have been developed which attempt to model the LHS, but they have almost exclusively developed these formulae for unidirectional flows.

As a result, there is no well-accepted theory to establish the relationship between the maximum combined shear stress τ_m and initiation of motion.

Therefore, we must resort to dimensional analysis. If we examine the forces on a particle, we find the following quantities regulate the dynamics: τ, g, D, ρ_f, ρ_p, ν.

We notice that the only terms with mass are the densities, which can be made to form the submerged specific gravity

$$R = \frac{\rho_p - \rho_f}{\rho_f} \tag{1}$$

which leaves two dimensionless variables to describe the system:

$$\tau^*_m = \frac{\tau_m}{R \rho g D} \quad \text{and} \quad Re_p = \frac{D\sqrt{\tau_m / \rho}}{\nu} \tag{2}$$

We postulate that the there exists some function *F* that relates the particle Reynolds number to the Shields stress at the point of initiation of motion. Mathematically,

$$\tau^*_{m_{crit}} = F\left(Re_p\right) \tag{3}$$

Shields (1936) was the first to propose such a relationship. Some of his data is plotted along with ‘recent’ data from Miller et al. (1977).

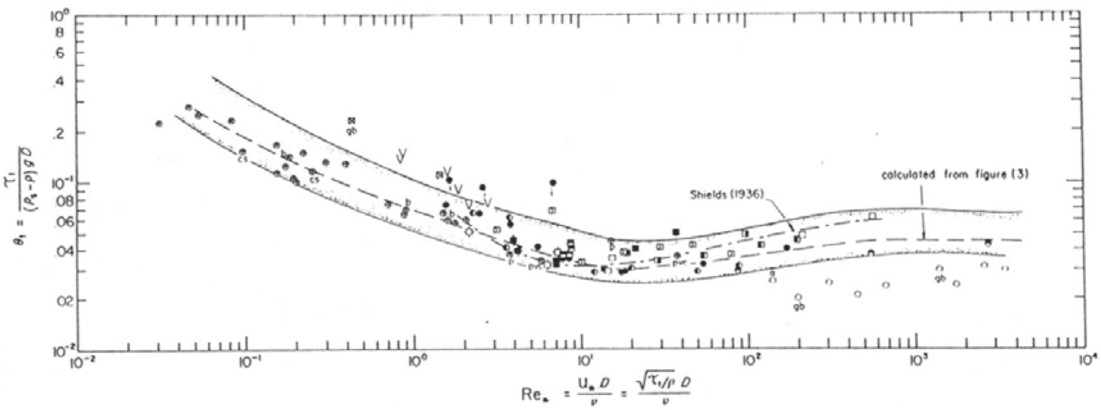


Fig. 2. The proposed modified Shields curve of θ_t versus Re_* based on additional carefully selected data. See Table 1 for identification of the symbols.

A problem with the Shields curve is that it is an implicit relation. It requires knowledge of the critical shear stress in order to calculate it. In other words, you generally have to iterate to find $\tau*_{m_{crit}}$.

Mixtures

Laboratory modelers typically use well-sorted materials to constrain *D*. However, natural sediments are often poorly sorted. Recently have there been efforts to account of the effects of the grain-size distribution. These formulations generally take the form –

$$\tau*_{crit_i} = \tau*_{crit_g}\left(\frac{D_i}{D_g}\right)^{\beta} \tag{4}$$

where D_i indicates the grain size in the ith size range in the mixture. D_g indicates the geometric mean grain size ($\overline{\Phi}$) of the 'surface layer'.

β takes a value of about 0.9 (Parker, 1990). Because beds typically **armor** themselves, a well-developed flow will often move sediment according to the shear stress of the largest material within the flow. This effect is often referred to as **hiding**.

Fine sediments

The Shields diagram is implicitly derived for sandy to gravelly beds. There are several assumptions that are often invalidated with fine sediment –

a. Fine sediment tends to be more poorly-sorted
b. Electrostatic forces become important (grain-grain cohesion and mineralogy dominate)
c. Definition of the 'bed' is more difficult
d. Turbulence and its structure strongly regulate movement.

As a result, initiation of motion is complex and dependent on several new factors. They are:

a. Sediment concentration
b. Degree of consolidation
c. Degree of flocculation

Entrainment into suspension

In many cases, we are interested in the rate of change of the bed height with time $d\eta/dt$. For most materials under the influence of waves, this is regulated primarily by suspension in the following relation

$$d\eta/dt = v_s\left(\overline{C_b} - E\right)/n_p \tag{5a}$$

where η is the height of the bed, $\overline{C_b}$ is the concentration of sediment at a height *b* above the bed, v_s is the settling velocity of the sediment and n_p is the porosity of bed.

This, of course, can be related to the flux of sediment into the water column

$$F_s = v_s\left(E - \overline{C_b}\right) \tag{5b}$$

Despite the appeal of a rigorous analysis, most models of *E* in wind-wave-induced environments have the Lickian form:

$$E = \alpha\left(\tau/\tau_{crit} - 1\right)^{\gamma} \tag{6}$$

where α and γ are empirical constants. I have dropped the subscript *m* – as it is implied in all of these formulae.

Ideally, if the data exist, you can fit Equation (6) to your field site. However, data of this quality are usually not available. Extension beyond the limits of your measured data is also questionable. As a result, you would like to rely on a generalized formulation.

The best work on wind-wave-induced resuspension has been done in lakes, not the ocean.

One popular (and theoretically sound) recent study is Luettich et al. (1990), who propose

$$E = v_s K \left(\frac{\tau - \tau_{crit}}{\tau_{ref}} \right)^n \qquad (7)$$

where the empirical parameters are related to one another as follows:

$$n = -0.67 \log_{10}(K) + 1.8 \qquad (8a)$$

and

$$\log_{10}(n) = -0.040 H_{crit} + 0.48 \qquad (8b)$$

where H_{crit} is the wave height in cm associated with the critical shear stress τ_{crit} predicted by linear wave theory. Luettich et al. (1990) used $\tau_{ref} = 0.072$ dyn cm^{-2}.

The model has successfully been used elsewhere (mostly in the Great Lakes). It can be used in combined wave-current settings, provided most of contribution to the shear stress is associated with waves (i.e., $\mu \to 0$). However, some have argued that wave-induced currents are always important in natural settings and cannot be ignored (Mei et al., 1997).

Bulk transport relations

In light of the absence of a generalized entrainment model, simpler relationships regulating bulk transport over regions have been proposed. Many of these are analogous to "stream power" arguments made for rivers.

Longshore (littoral) transport

Like the stream power arguments Ralph Bagnold pioneered in fluvial settings, we can assign a **wave power**

$$P_l = (ECn)_b \sin\alpha_b \cos\alpha_b \tag{9}$$

where the subscript b denotes that all of the quantities be calculated at the breaker line.

The energy consumed by the wave power will be expressed as an "immersed-weight transport rate"

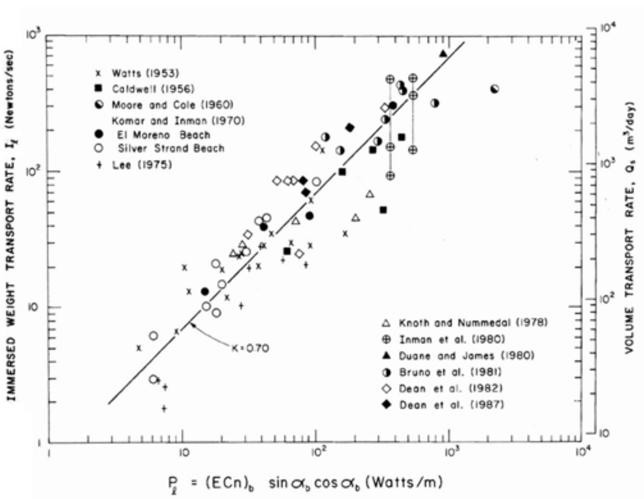
$$I_l = (\rho_s - \rho_f)(1 - n_p) g Q_l \tag{10}$$

where Q_l is the volumetric transport rate of sediment, ρ_f is the density of the water and ρ_s is the density of the sediment.

Bagnold (1963, 1966), like in the fluvial case, saw the relationship between I_l and P_l as an inefficient process, in which an efficiency K is needed. Mathematically, this results in

$$I_l = KP_l = K(ECn)_b \sin\alpha_b \cos\alpha_b \tag{11}$$

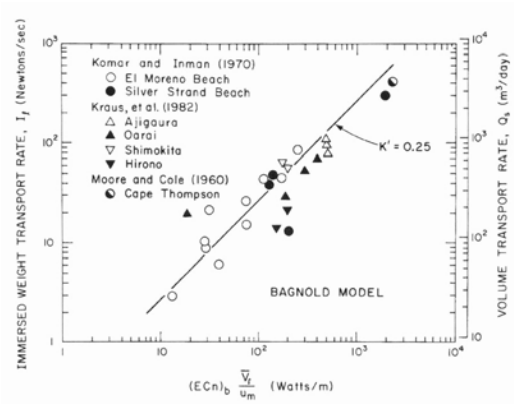
Komar recommends a value of $K = 0.7$.



Bagnold (1963, 1966) also suggest an alternative

$$I_l = K' \left(\frac{ECn}{u_{bm}} \right)_b v_l \tag{12}$$

where the subscript *b* denotes the quantities be evaluated at the breaker line. The longshore velocity v_l is usually evaluated at the **mid-surf position**. Equation (12) may make more sense, given the type of data available at a particular site. Komar suggests a value of $K' = 0.7$.



Of course, there have been laboratory experiments, which have sought to tie wave characteristics more closely to the physics of sediment transport – and overcome some of the biases present in existing field measurements.

The recognized state-of-the-art model is Kamphuis (1990), who suggests

$$\frac{Q_l}{\rho H_{bs} / T} = 0.0013 (\tan \beta)^{0.75} \left(\frac{H_{bs}}{L_\infty} \right)^{-1.25} \tag{13}$$

As with any laboratory experiment, the relationship possesses many possible scale effects. Kamphuis (1990), as discussed by Komar, is rife with inconsistencies to natural data.

Wave (marine) environments are more difficult to model in the laboratory as compared to their fluvial (river) cohorts.

Cross-shore variability in longshore transport

Komar discusses several models of longshore transport. However, considering how poorly constrained the cross-shore variability in longshore currents are, these models are only descriptive at best.

From these analyses and measurements it is clear that the peak in alongshore sediment transport occurs seaward of the peak in the velocity (typically around the mid-surf position).

Cross-shore transport

There are three primary ways sediment can be moved cross-shore: 1) strong (rip) currents associated with nearshore circulation, 2) by asymmetric wave motions in the wave boundary layer and 3) by sediment-induced gravity currents.

Ironically, the best understood mechanism is the first. This is in spite of its relatively young age compared to the last two possibilities (both of these mechanisms have been known and studied for nearly a century).

The last two possibilities are strongly controlled by the local geology and sediment supply. Fine material (even just a fine tail) strongly affects the bottom boundary layer – impeding boundary layer energy and collecting negative buoyancy.

The section in Komar's book dedicated to cross-shore transport is not particularly helpful and in some instances incorrect or out-of-date.