*Ann. Rev. Earth Planet. Sci. 1986. 14: 113–47*


# OCCURRENCE AND FORMATION OF WATER-LAID PLACERS

*Rudy Slingerland*

Department of Geosciences, The Pennsylvania State University, University Park, Pennsylvania 16802

*Norman D. Smith*

Department of Geological Sciences, University of Illinois at Chicago, Chicago, Illinois 60680

## INTRODUCTION

A placer is a deposit of residual or detrital mineral grains in which a valuable mineral has been concentrated by a mechanical agent. The agent is usually running water, and the valuable mineral is usually denser than quartz (for example, gold, diamond, cassiterite, ilmenite, or chromite). It is difficult to overemphasize the importance of placers as sources of mineral wealth; the Witwatersrand (South Africa) paleoplacers, discovered in 1886, alone have provided over half of all the gold ever mined in the world (Pretorius 1976), and alluvial placer deposits of cassiterite in southeast Asia presently are the world's major source of tin (Toh 1978). The earliest evidence of placer mining comes from stone carvings in Egypt dated at 2500 B.C. (MacDonald 1983), and as early as 7 B.C., Strabo pointed out that in Turdetania (Spain) more gold was procured from washing sand in the rivers than by digging in the mines.

Given the economic importance of water-laid placers and the long history of their exploitation, it is remarkable that they have received so little attention in sedimentary research. Mining and exploration practices for the most part have been guided by empirical rules of thumb that are locally useful but lack any firm physical basis of understanding [see Bateman



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(1950) and MacDonald (1983) for summaries]. For example, MacDonald (1983, p. 145) states that "on normally stable beaches the highest concentrations [of heavy minerals] are found where the pounding of waves has been most intensive . . . ," although the causes of this association, even if true, remain obscure. It is significant, then, that in the last decade a body of knowledge on the occurrence and genesis of placers has grown out of both the physical principles of hydrodynamics and sediment transport and an improved understanding of the sedimentary environments of deposition. To be sure, whether or not an economic placer deposit forms may depend ultimately on such external factors as the stage of stream evolution (Tuck 1968, Kartashov 1971, Schumm 1977, Adams et al 1978), tectonic history (Henley & Adams 1979, Sigov et al 1972), local geology and physiography (Jenkins 1964), or climate (Krook 1968). But common to the origins of all water-laid placers are concentrating mechanisms that involve interactions among the fluid, sediment bed, and transported particles, and it is these mechanisms for sorting dense sediment grains from light grains that we emphasize in this review. Unfortunately, our coverage is not exhaustive. Russian papers are numerous in this field, and only a few translations are considered here.

## OCCURRENCES OF WATER-LAID PLACERS

Water-laid placers occur in a variety of geomorphological sites (observed or interpreted) and over a wide range of physical scales. Undoubtedly the most common and important group of economic placer deposits are those formed by streams, with beaches and shallow nearshore areas probably second in importance. Other settings, by comparison, are relatively minor, either in modern environments or as ancient paleoplacers.

Table 1 lists sites of heavy-mineral concentrations described from both modern and ancient examples and classifies them according to their spatial scales (Smith & Minter 1980, p. 1; Slingerland 1984, p. 138). Large-scale concentrations (order $10^4$ m) occur on regional or system-wide scales as products of long-term interactions among time-averaged flow variables, available heavy minerals, and substrate characteristics. The intermediate scale (order $10^2$ m) refers to concentrations associated with major depositional or erosional topography within the sediment-transporting system. In fluvial settings, these are commonly bars, short channel segments, riffles, and the like. Small-scale concentrations (order $10^0$ m) occur at the sediment-bed scale and are commonly manifested as heavy-mineral-rich laminations in sequences of stratification. Sorting associated with the formation and migration of bedforms is a dominant cause of small-scale segregations. These scales are clearly hierarchical in that smaller scales

Table 1 Observed sites of water-laid placers

| Sites | References |
|---|---|
| Large scale ($10^4$ m) | |
| Bands parallel to depositional strike | Minter 1970, 1978, Sestini 1973, McGowen & Groat 1971 |
| Heads of wet alluvial fans | Schumm 1977 |
| Points of abrupt valley widening | Kuzvart & Bohmer 1978, Crampton 1937 |
| Points of exit of highland rivers onto a plain | Toh 1978 |
| Regional unconformities | Minter 1976, 1978 |
| Strand-line deposits | Nelson & Hopkins 1972, Komar & Wang 1984, Eliseev 1981 |
| Incised channelways | Minter 1978, Yeend 1974, Buck 1983 |
| Pediment mantles | Krapez 1985 |
| Intermediate scale ($10^2$ m) | |
| Concave sides of channel bends | Kuzvart & Bohmer 1978, Crampton 1937 |
| Convex banks of channel bends | Kuzvart & Bohmer 1978 |
| Heads of midchannel bars | Toh 1978, Smith & Minter 1980, Kartashov 1971, Boggs & Baldwin 1970 |
| Point bars with suction eddies | Toh 1978, Bateman 1950 |
| Scour holes, especially at tributary confluences | Kuzvart & Bohmer 1978, Mosley & Schumm 1977 |
| Inner bedrock channels and false bedrock | Schumm 1977, Kuzvart & Bohmer 1978, Adams et al 1978 |
| Bedrock riffles | Cheny & Patton 1967, Toh 1978 |
| Constricted channels between banks and bankward-migrating bars | Smith & Minter 1980, Smith & Beukes 1983 |
| Beach swash zones | Stapor 1973, Reimnitz & Plafker 1976, Kogan et al 1975 |
| Small scale ($10^0$ m) | |
| Scoured bases of trough cross-strata sets | Toh 1978, McGowen & Groat 1971, Smith & Minter 1980, Buck 1983 |
| Winnowed tops of gravel bars | Toh 1978, McGowen & Groat 1971 |
| Thin ripple-form accumulations | Brady & Jobson 1973 |
| Dune crests | Brady & Jobson 1973 |
| Dune foresets | Brady & Jobson 1973, McGowen & Groat 1971, Buck 1983 |
| Plane parallel laminae | Slingerland 1977, Clifton 1969, Buck 1983, Stavrakis 1980 |
| Leeward side of obstacles | Lindgren 1911 |
| Beach berms | Stapor 1973 |

are superimposed on larger ones; thus, a few heavy minerals might concentrate on a small scale over a very short time in response to some sorting event, but increasingly larger scales of concentrations require successively greater areas over which the sorting mechanism is applied. For example, a strand-line placer (large scale) may be dominated by patchy concentrations in the beach swash zone (intermediate scale); these concentrations in turn are composed of segregated heavy-mineral-rich laminations (small scale). A basic understanding of placer formation, therefore, requires knowledge of sorting arising from the finer scales of fluid-sediment interactions; such knowledge is only partial at this time.

## CONDITIONS FOR PLACER DEVELOPMENT

The various placer sites listed in Table 1 contain five factors common to all members. First, in the parlance of mining technology, each site is a natural geomorphological "dressing mill." The mill may be the toe of a dune avalanche face or a berm crest on a beach, but in each case it receives "pulp" (that is, granule- to silt-sized, discrete heavy and light mineral grains mixed with water) and remains relatively fixed in time and space as it sizes and separates the pulp. Second, heavy minerals are present in the pulp and have a size distribution that is proper for the natural dressing mill. Most alluvium or natural pulp contains some heavy minerals, but usually in concentrations of less than 1%. Gemstones and gold are much rarer, of course, and are linked to particular plate tectonic settings (Henley & Adams 1979). The initial size distributions of heavy and light minerals in various source terranes are still not known with any accuracy. Given the crystal sizes of minerals in igneous and metamorphic rocks (Feniak 1944) and the increased susceptibility of some heavy minerals to comminution, the global average grain size of heavy minerals in alluvium is probably less than that of light minerals, even before hydraulic sorting has occurred (Rittenhouse 1943, Van Andel 1959, Briggs 1965, Stapor 1973). This assumption is made in the subsequent discussions. Third, in each case the pulp is fed at a proper rate for the mill size. Fourth, and especially critical, each dressing mill contains the proper combination and sequence of hydraulic sorting mechanisms for separating the pulp by size and density. Fifth, the geometry of the mill circuits is such that tailings are properly disposed (transported away) and ore concentrate is quasi-permanently stored.

A tributary junction of a stream is a good example of an intermediate-scale natural dressing mill (Schumm 1977). It processes pulp from a tributary, stores the ore concentrate in deep scour holes, and disposes of tailings by sediment transport due to flows of the main channel. The pulp rate is governed by a feedback loop from the mill through a base-level control on the amount of upstream erosion. The sorting mechanisms created by the local flow geometry are the sediment entrainment and transport processes discussed in what follows.

## HYDRAULIC SORTING MECHANISMS—THE SINE QUA NON

However favorable other factors may be, without mechanisms for sorting dense sediment grains from light grains, there will be no water-laid placer deposit. Sorting of a heterogeneous size-density pulp is accomplished in natural mill circuits by an alternation of at least four mechanisms, some sorting more by size and some by density. The sorting may be subdivided by scale into two types: local and progressive (Brush 1965, Rana et al 1973, Deigaard & Fredsøe 1978). Progressive sorting occurs by the cumulative effects of local sorting and along-flow changes in competency and capacity.

The mechanisms of local sorting are the following (Slingerland 1984): (*a*) the free or hindered settling of grains, usually in turbulent water, (*b*) the entrainment of grains from a granular bed by flowing water, (*c*) the transport of grains by flowing water, and (*d*) the shearing of grains in a moving granular dispersion. The categories are not mutually exclusive—shear sorting may play a role in transport sorting, for example—but we find this organization useful at our present level of knowledge.

Before discussing each mechanism in detail, we briefly highlight some pertinent principles of fluid mechanics and sediment entrainment. Middleton & Southard (1977) and Yalin (1977) provide excellent, more in-depth reviews. Flows in natural stream channels are either predominantly or wholly turbulent (i.e. they are characterized by random velocity fluctuations and flow pathlines that are strongly three dimensional). The gross character of the flow can be visualized by considering a vertical cross section in the downstream direction with a planar channel bottom (Figure 1). Three intergrading flow zones can be identified for turbulent flows above a smooth boundary (Hinze 1975): (*a*) Immediately above the channel bottom is the *viscous sublayer*, which contains turbulent fluctuations but is dominated by viscous rather than turbulent transfer of momentum across fluid layers; it rarely exceeds a few millimeters in thickness. (*b*) A thin *turbulence-generation layer* (or "buffer layer") is located just above the viscous sublayer. Here the shear stresses are very high, and small but strong turbulent eddies are generated and carried outward above the boundary or downward into the viscous sublayer. (*c*) The *outer* (or *core*) *region* occupies the remainder of the flow boundary layer; in most natural streams, this zone comprises most of the flow depth and the highest mean velocities.

In hydrodynamically "rough" bottoms, coarse sediment particles project upward sufficiently to disrupt flow structure near the bed; the viscous sublayer is destroyed, and the turbulence-generation layer extends to the bottom. This situation occurs when the height of the roughness elements, $K$ [taken by Einstein (1950) as the 65th percentile of the bottom grain size distribution] exceeds the "potential" thickness of the viscous sublayer [i.e. its thickness if the boundary were smooth under existing flow conditions (Figure 1)].

The curved profile of upward-increasing velocity (Figure 1), typical of open-channel flow, is caused by bottom frictional drag, which is transferred upward by viscous (in laminar flow) and inertial (turbulent eddies) momentum exchanges between successive fluid layers. If we assume a straight channel segment in which neither depth nor average velocity changes downstream (i.e. "uniform flow"), the force exerted by the gravity-driven flow on the channel boundary (bottom and sides) is balanced by the frictional resistance of the boundary, because the assumed flow neither accelerates nor decelerates. The temporal mean boundary fluid force, called the tractive or boundary shear stress $\bar{\tau}_0$, is given by

$$\bar{\tau}_0 = \rho_f g R S, \tag{1}$$

in which $\rho_f$ is the fluid density, $g$ the acceleration due to gravity, $S$ the slope of the stream bed and water surface (equal in uniform flow), and $R$ the hydraulic radius. For natural streams in which the width greatly exceeds

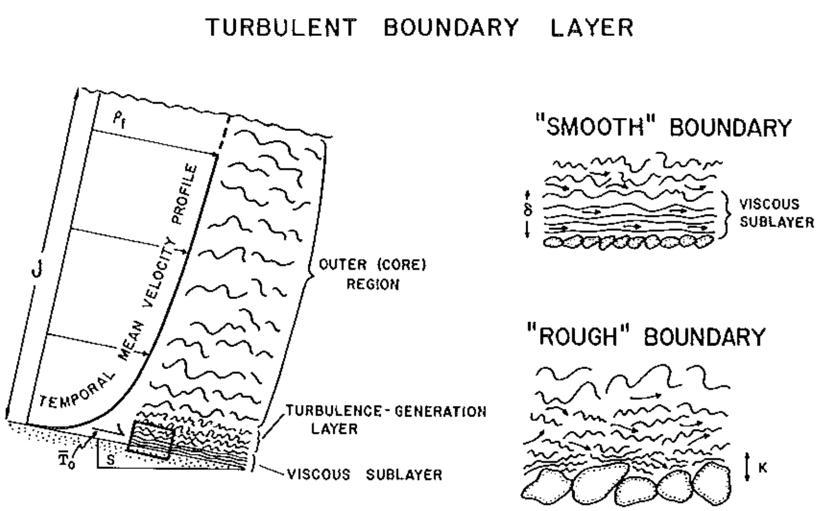


*Figure 1* Internal structure of natural turbulent flows, where $J$ is the flow depth, $\rho_f$ the fluid density, $\bar{\tau}_0$ the temporal mean boundary shear stress, $S$ the bed surface slope, $\delta$ the thickness of the viscous sublayer, and $K$ the height of roughness elements. See text for discussion.

the depth $J$, it follows that $R \sim J$ and thus that

$$\bar{\tau}_0 = \rho_f g J S. \tag{2}$$

The instantaneous shear stress at the bed follows a positive skewed distribution, with a coefficient of variation equal to 0.4 (Grass 1983); thus, grains may experience shear stresses up to twice the temporal mean given in Equations (1) and (2).

For certain problems in fluid mechanics, a parameter defined as

$$U_* = \sqrt{\frac{\bar{\tau}_0}{\rho_f}} = \sqrt{gJS} \tag{3}$$

is used, in which $U_*$ is termed the shear velocity, or friction velocity. Although $U_*$ has the dimension of velocity, it need be considered only as a surrogate or "convenience" variable for tractive shear stress. The shear velocity cannot be measured directly (such as with a current meter) and is always much smaller than the average velocity.

The presence, and thickness, of the viscous sublayer is important to bottom sediment movement because this sublayer affects the nature and distribution of fluid forces (viscous and pressure) acting on the grains. For smooth boundaries, the thickness of the sublayer, $\delta$, depends on shear velocity and viscosity as

$$\delta = \frac{c\nu}{U_*}, \tag{4}$$

where $c$ is a constant. Since protruding grains of sufficient size will destroy the viscous sublayer, it is reasonable to think that the ratio of bottom roughness size to sublayer thickness, $K/\delta$, may serve to define hydrodynamically rough and smooth boundaries. Substituting from (4), we have

$$\frac{K}{\delta} = \frac{U_* K}{c\nu} = \text{constant} = R_*, \tag{5}$$

where the dimensionless quantity $U_* K/\nu$ is termed the boundary Reynolds number $R_*$. Earlier workers (see Inman 1949) considered that the value $R_* = 3.5$ distinguished smooth from transitional boundaries, but $R_* = 5$ is the more commonly accepted value today. The value of $R_*$, then, reflects the degree to which grains on the bed project into the turbulent zone of the boundary layer, and we should expect the distribution of fluid forces acting on grains to be a function of that value. For $R_* > 70$, the wall is considered to be fully rough, with the range $5 < R_* < 70$ representing a transition in which turbulence only periodically disrupts the viscous sublayer and impinges directly on the grains.

preciable, say greater than 5%, their fall is hindered by grain-grain interactions and an upward counterflow of the suspending fluid. In monodisperse systems, it is generally agreed that a grain's constant terminal settling velocity $w_\infty$ is retarded according to the expression $w/w_\infty = (1-C)^n$, where $w$ is the hindered settling velocity, $C$ the volumetric grain concentration, and $n$ an exponent equal to $4.4\,\mathrm{Re}^{-0.1}$ in the range $200 < \mathrm{Re} < 500$ (Richardson & Zaki 1954). In polydisperse systems of mixed sizes and densities, no general theory yet exists, but the experimental studies of Davies (1968), Lockett & Al-Habbooby (1974), Mirza & Richardson (1979), and especially Richardson & Meikle (1961) have shown that the above equation applies to individual species in a mixture if $C$ is defined as the total concentration of all species present.

How does free settling of grains aid in the formation of placers? If the settling velocity distributions of the heavy and light minerals in a pulp are equal [i.e. hydraulically equivalent in Rubey's (1933, 1938) sense], then free settling by itself will not sort the grains. Many heuristic explanations of placer development and heavy-mineral occurrences low in an alluvial fill assume that the heavy minerals settle faster than the light minerals (MacDonald 1983); these explanations do not consider that the starting heavy-mineral size distributions probably have smaller means and therefore are potentially hydraulically equivalent to the light minerals. Free settling of hydraulically equivalent distributions is important, however (as is discussed later), because it produces deposits that can be subsequently sorted by size. If the pulp contains heavy and light minerals with unequal settling velocity distributions, then its free settling in a stationary fluid will sort the grains and produce a deposit vertically layered by density. If the free settling occurs in a unidirectional flow, then a deposit laterally segregated by density will be formed. Finally, if the pulp is the suspended load of a turbulent flow, grains will be fractionated into different elevations above the bed (Rouse 1950, Brush 1965, Middleton & Southard 1977, p. 6.27) such that at any specified elevation, the ratio of concentration $C$ to the concentration at some reference level $C_a$ for heavy (h) versus light (l) mineral grains is

$$\left(\frac{C}{C_a}\right)_h = \left(\frac{C}{C_a}\right)_l^{w_h/w_l}, \tag{7}$$

where $w_h$ and $w_l$ are the settling velocities of the heavy and light grains, respectively (Slingerland 1984). If the mean settling velocities of the heavy grains are smaller than those of the light grains, perhaps as a result of local size deficiencies, then higher levels of the flow will be relatively enriched in suspended heavy grains, which can then be carried to higher elevations of the floodplain. Such a mechanism is suggested by the recent work of Nami

(1983), who shows in a small section of the Witwatersrand Carbon Leader Reef that interchannel highs contain greater concentrations of placer gold than thicker contiguous channel deposits.

How does hindered settling of grains aid in the formation of placers? As an example, consider a mixture used by Richardson & Meikle (1961) consisting of equal volumes of two species of sand-sized spheres, one of density 2.9 g $cm^{-3}$ and diameter 0.071 mm and the other of density 1.04 g $cm^{-3}$ and diameter 0.382 mm such that they have equal free settling velocities. A grain density of 1.04 is unrealistic for natural minerals, but the results should still be applicable to common mixtures of light and heavy grains. If the total concentration $C$ is less than 8%, the two species settle with equal (but reduced) velocities and produce a single mixed layer. If $C$ is between 8 and 10%, settling produces three layers of sediment, the lower consisting solely of the denser grains, the upper consisting solely of the lighter grains, and the intermediate consisting of a mixture. At concentrations above 10%, settling produces two layers, each completely segregated by size and density. The results were explained by Richardson & Meikle in terms of the buoyant forces on the large light grains created by the mixture of fluid and small dense grains. This phenomenon may be the principal source of mineral sorting in jigs and hydraulic classifiers, and it deserves more study. In natural dressing mills, the necessary conditions for its operation might be realized in decelerating overwash flows on beaches or points of flow expansion in streams where sediment drops quickly out of suspension. The resulting deposits would be laminae or beds with heavy-mineral-enriched bases.

## *Entrainment of Grains*

Entrainment is the dislocation of grains from a granular bed and their initial movement by a superimposed fluid flow. In realistic situations, the grains possess differing sizes, densities, and shapes, and the flow is unidirectional, nonuniform, unsteady, and turbulent. It is generally accepted that under these conditions the probability of a grain's entrainment increases as the mean values and variances of the fluid forces increase over the gravity and frictional forces holding the grain in place [see Yalin (1977) for a review]. Sorting occurs and placers may be formed because the forces depend upon a grain's size, shape, and density. The phenomenon is a stochastic one because (*a*) the magnitudes of the fluid forces in a turbulent fluid comprise a Gaussian or positively skewed distribution in time (Grass 1983) and (*b*) the detaining forces vary with the local grain geometries of the bed. For these reasons it is convenient to define a dimensionless parameter $N$ equal to $nD/U_*$, where $n$ is the number of grains of diameter $D$ in motion per unit area per unit time, and $U_*$ is the fluid shear velocity (Yalin 1977).

Then the "critical" fluid, grain, and bed values at the threshold of motion may be evaluated for any arbitrarily small $N$; Yalin suggests $N = 10^{-6}$.

The theoretical prediction of critical conditions proceeds from a torque balance on a grain in the bed (White 1940, Everts 1973, Yalin 1977, Middleton & Southard 1977, Slingerland 1977; see Figure 3). Most researchers (see Graf 1971) simplify the situation by assuming that the bed is horizontal, planar, and cohesionless, that no grains are yet moving in the flow, and that the grains are equant. For a grain to rotate about a pivot point A (Figure 3), a moment balance shows that the ratio of the fluid forces F to the gravity forces G must be

$$\frac{F}{G} \geq \frac{a_g \sin \alpha}{a_f \cos (\alpha - \zeta)}, \tag{8}$$

where $a_g$ and $a_f$ are the moment arms about A, $\alpha$ is the reactive angle, $\zeta$ is the angle between the fluid force vector and the horizontal, and the point of application of the fluid force is assumed to be along the normal to the pivot

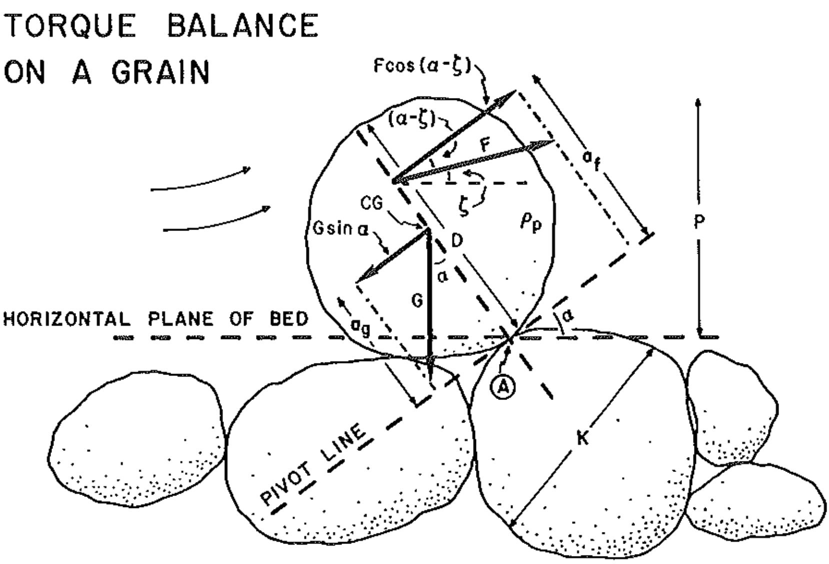


*Figure 3* Definition diagram for calculating the initiation of grain motion on a horizontal bed. A subspherical grain of diameter $D$ and density $\rho_p$, protruding above the bed a distance $P$, must pivot about point A on a downflow grain of diameter $K$. A fluid force vector $F$, representing both drag and lift forces, acts at a distance $a_f$ away from a pivot line through A and at an angle $\zeta$ from the horizontal. A grain weight vector $G$ acts through the grain center of gravity CG at a distance $a_g$ away from the pivot line. At the moment of entrainment, the fluid torque must be greater than the resisting torque.

line. The magnitude of the fluid force vector $F$ and its orientation $\zeta$ are determined by the magnitudes of the drag and lift forces and their points of application on the grain, all inadequately known. Some derivations (Slingerland 1977, Sundborg 1956) relate the drag and lift forces to the square of the local mean flow velocity near the top of the grain using lift and drag force equations, whereas others (Yalin 1977) relate them to the temporal mean boundary shear stress $\bar{\tau}_0$ or its surrogate, the shear velocity $U_*$. Because the lift and drag forces may be related to $U_*$ and the boundary Reynolds number $R_*$ (Einstein & El-Samni 1949, Coleman & Ellis 1976), and also because of the difficulty of defining the velocity at the top of the grain, it is probably preferable to express the resultant fluid force magnitude in the latter manner, such that

$$F = f(bD^2, \bar{\tau}_0, R_*), \tag{9}$$

where $b$ is a proportionality factor relating $D^2$ to the true area of the grain on which the fluid forces act, $D$ is the diameter of the grain, and $\bar{\tau}_0$ and $R_*$ are defined in Equations (1) and (5), respectively. The gravity force $G$ may be written as the grain's submerged weight,

$$G = cD^3(\rho_p - \rho_f)g, \tag{10}$$

where $c$ accounts for the grain's nonspherical shape. The moment arm $a_g$ is equal to the grain radius $D/2$ if the grain is a sphere, or more generally to $cD$. The moment arm $a_f$, proportionality factor $b$, and angle $\zeta$ vary with the structure of the fluid boundary layer (measured by $R_*$) and the protrusion of the grain above the mean bed elevation (measured by $P/D$; Abbott 1974) and must be evaluated experimentally. The reactive angle $\alpha$ is the angle of repose of a single grain on a fixed bed and decreases as $D/K$, $D$, sphericity, and roundness increase (Miller & Byrne 1966, Carrigy 1970, Luque 1974, Z. Li & P. D. Komar; see Refs. Added in Proof).

Defining $\bar{\tau}_c$ as the mean critical boundary shear stress at the threshold of motion and substituting Equations (9) and (10) into (8) while considering the above yields

$$\frac{\bar{\tau}_c}{(\rho_p - \rho_f)gD} \geq f(D/K, P/D, R_*). \tag{11}$$

The left-hand side is called $\theta_c$, the critical dimensionless Shields parameter (Shields 1936), and is the ratio of the shear stress exerted by the fluid on the bed to the weight of a potentially entrainable grain layer over a unit area.

In the simplest case, where well-sorted sediments are entrained and transported over a bed of equal-sized particles, $\theta_c$ is solely a function of the boundary Reynolds number. A graph of $\theta_c$ versus $R_*$ is difficult to interpret, however, because both parameters contain the fluid shear stress. To circumvent this difficulty, Yalin (1977) divided $R_*^2$ by $\theta_c$ and obtained a dimensionless parameter $\Xi$, whose square root is the Yalin parameter. At Yalin numbers above 40, experimental data from numerous studies using differing grain and fluid densities define the relationship quite well (Figure 4; Miller et al 1977, Yalin & Karahan 1979). The variance of data at any one boundary Reynolds number is in part due to the various definitions of $N$ (that is, of how many grains must be moving at the threshold of motion). At Yalin numbers between 4 and 40, Collins & Rigler (1982) found that $\theta_c$ was overestimated for grains of density from 4 to 7 g cm$^{-3}$. In fact, for grain settling velocities $w_\infty$ less than 10 cm s$^{-1}$ and a wide range of densities, the critical boundary shear stress in their experiments was solely a function of grain settling velocity and independent of $R_*$, such that

$$\bar{\tau}_c = 1.24\, w_\infty^{1/3}. \tag{12}$$

This is not a theoretically predictable relationship (P. D. Komar & K. E. Clemens; see Refs. Added in Proof), and until more experiments are completed, it remains a troublesome complication.

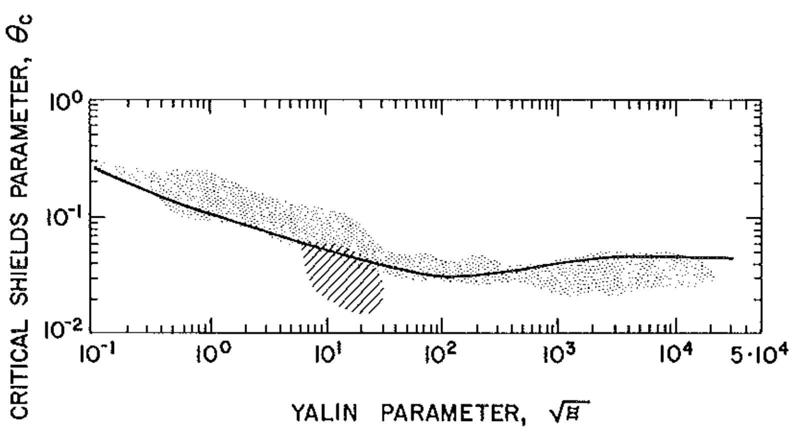


*Figure 4* Standard threshold conditions for entrainment of grains of uniform size (redrawn from Miller et al 1977). The parameters $\theta_c$ and $\sqrt{\Xi}$ are defined by

$$\theta_c = \bar{\tau}_{0c}/[(\rho_p - \rho_f)gD],$$

$$\sqrt{\Xi} = [(\rho_p - \rho_f)gD^3/(\rho_f \nu^2)]^{1/2},$$

where $\bar{\tau}_{0c}$ is the critical boundary shear stress, $\rho_p$ is the grain density, $\rho_f$ is the fluid density, $g$ is the gravitational acceleration, $D$ is the grain diameter, and $\nu$ is the kinematic viscosity of the fluid. The stippled pattern denotes the data field used by Miller et al in defining their preferred curve (shown as a black line). The diagonal pattern denotes the data field of Collins & Rigler (1982).

How does entrainment of well sized-sorted sediments lead to the separation of grains into populations of differing densities? Recasting the curve in Figure 4 as a plot of critical shear stress against grain diameter (Figure 5) yields simple monotonically rising curves; curves for different densities are of similar form but progressively displaced toward higher values of shear stress. As intuitively expected, less dense grains can be selectively winnowed from a population of equal-sized but denser grains, leaving a denser lag. Ljunggren & Sundborg (1968), McQuivey & Keefer (1969), and Grigg & Rathbun (1969) were among the first to analyze heavy-mineral entrainment using Shields criteria of this type. It must be remembered, however, that this analysis applies only to well-sorted sediments on a plane bed. It is not possible to predict from Figure 5 which sizes of different density minerals will be entrained off a bed together, because the presence of different sizes violates the experimental conditions under which the curve was obtained. Certain thin, planar, heavy-mineral-rich laminae in fluvial deposits such as described by Lucchitta & Suneson (1981) and Stavrakis (1980) might be explained in this manner.

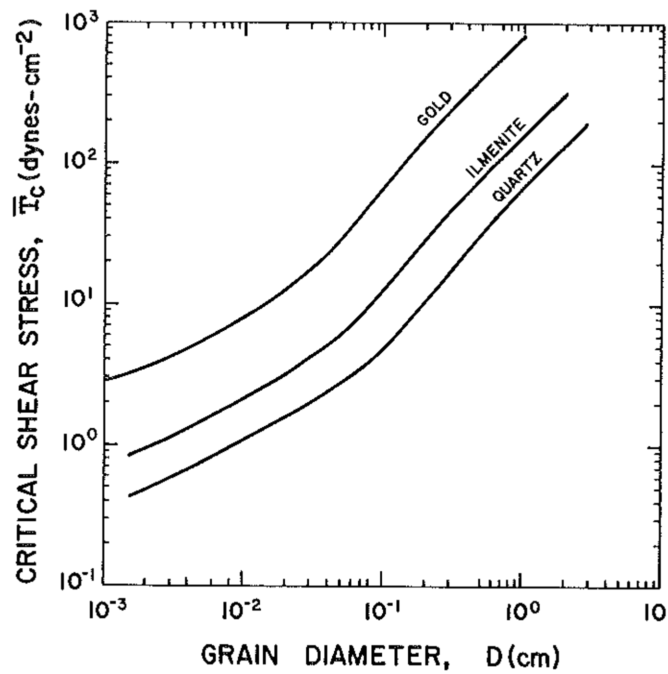


*Figure 5* The effect of grain density on the critical boundary shear stress. Curves are derived from Figure 4 using densities for quartz, ilmenite, and gold of 2.65, 4.70, and 19.3 g cm$^{-3}$, respectively.

In the more general case, where grains of varying sizes and densities compose the bed, all three variables on the right side of Equation (11) are important. Larger grains protruding higher into the flow present a greater surface area for drag and lift forces, experience greater instantaneous turbulent shear stresses, and have smaller reactive angles $\alpha$. Smaller grains are either sheltered from the flow or experience increased turbulence from the wakes of larger grains. A curious result is that coarser grains may be entrained at shear stresses lower than those for finer grains (Gilbert 1914, Meland & Norrman 1966, 1969, Brady & Jobson 1973, Saks & Gavshina 1975, Slingerland 1977, Day 1980, Raudkivi & Ettema 1982); thus a sorting mechanism based more on size exists to complement the sorting mechanisms based more on density.

In the simplified case of a binary mixture where particles of diameter $D$ rest in a bed of particles of diameter $K$ and protrude to varying heights $P$ above the mean bed level, the appropriate Shields parameter in Equation (11) for the superjacent particle (labeled $\theta_{ci}$ to separate it from its value for a level, uniformly sized bed) can be calculated from an equation due to Slingerland (1977, Equation 12). Written in terms of shear stress for a horizontal bed, the equation is

$$\theta_{ci} = \frac{4}{3} \frac{(\beta_1 \beta_2)^2 \tan \alpha}{C_d}. \tag{13}$$

Here, $\beta_1$ is a turbulent velocity fluctuation coefficient, $\beta_2$ is a coefficient accounting for the point of application of fluid forces and angle $\zeta$ in Figure 3, and $C_d$ is the coefficient of drag appropriate for the grain as it rests in the bed. All are functions of the boundary Reynolds number $R_*$ and the protrusion distance $P/D$.

The reactive angle $\alpha$ is determined by the protrusion distance $P/D$ and the size of the grain relative to the underlying grains, $D/K$ (Figure 3). In the case where a grain protrudes above an adjacent downflow grain of equal size, the geometry is such that

$$\alpha = \arccos P/D. \tag{14}$$

In the case where a grain of size $D$ rests on grains of size $K$, Z. Li & P. D. Komar (see Refs. Added in Proof) have shown that

$$\alpha = e(D/K)^{-f}, \tag{15}$$

where $e$ varies from 35 to 70 and $f$ varies from 0.30 to 0.75. For nearshore marine sands, Miller & Byrne (1966) give $e = 61.5$ and $f = 0.3$; for gravel-sized spheres undergoing grain-top rotation, Z. Li & P. D. Komar (see Refs. Added in Proof) give $e = 36.3$ and $f = 0.72$. For river gravels where $K$ is taken as equal to the mean grain size $\bar{D}$, P. D. Komar & Z. Li (see

Refs. Added in Proof) give $e = 35$ and $f = 0.9$ for $D_i/\bar{D} > 1$ and $e = 35$ and $f = 0.6$ for $D_i/\bar{D} < 1$.

The influence of protrusion distance on $\theta_{ci}$ is seen in Figure 6, in which are plotted the data of Fenton & Abbott (1977) collected in a fully turbulent boundary layer. In their experiment, particles were slowly pushed up into the flow through adjacent fixed grains until swept away. At the moment of entrainment, the relative distance of protrusion above the local bed level was $P/D$. Also shown in Figure 6 is the critical Shields parameter $\theta_c = 0.045$ for a level, uniformly sized, hydrodynamically rough bed. It can be seen from the figure that particles that protrude above the top of the bed a distance less than one third to one quarter their diameter possess $\theta_{ci}$ larger than 0.045, whereas particles protruding farther into the flow possess lesser $\theta_{ci}$. The causes of this deviation are not well understood, but most certainly

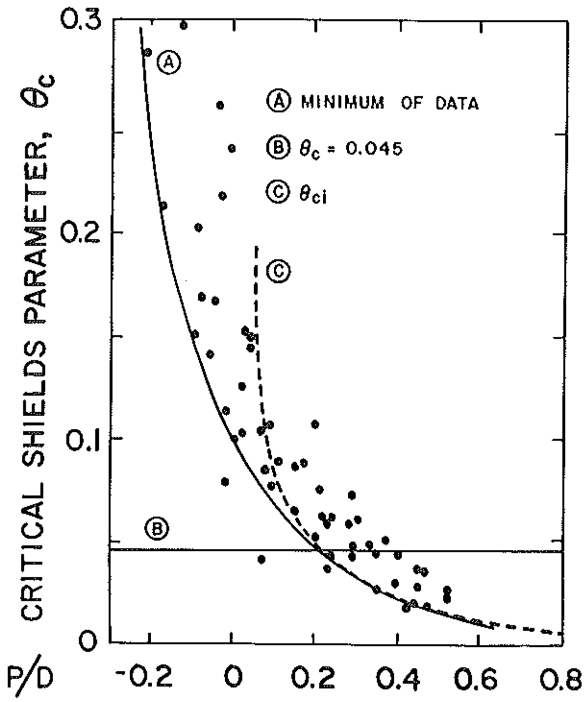


*Figure 6* The effects of relative grain protrusion $P/D$ on the critical Shields parameter. Data points are from flume experiments by Fenton & Abbott (1977, experiment B) at $\sqrt{\Xi} = 500$ and define a minimum curve A (excluding outliers). Line B is the conventional critical Shields criterion from Figure 4 for these conditions. Curve C is a plot of Equation (13) in the text. Grains can be less or more easily entrained depending upon their relative protrusion above the bed.

they include a reduction in the reactive angle $\alpha$, an increase in the surface area of the grain upon which the fluid shear stress acts, and an increase in the intensity of turbulence as grains protrude higher into the flow. To explain the relative contributions of each, it may be assumed as a first approximation that for a fully rough bed the variation due to turbulence and surface area is much less than that of tan $\alpha$. Then Equation (13) can be plotted in Figure 6 by substituting Equation (14) for $\alpha$ and evaluating the remaining terms as a constant, calculated to yield $\theta_{ci} = 0.045$ when $P/D = 0.2$. Although the data reflect a unique grain geometry and Equation (13) fits only moderately well, the trends and relative magnitudes of $\theta_{ci}$ seem indisputable. Much of the decrease in the critical Shields parameter with increasing protrusion can be ascribed to a decrease in the reactive angle.

Similar theoretical results have been obtained by Komar & Wang (1984) and P. D. Komar & Z. Li (see Refs. Added in Proof) for the variation in entrainment threshold due to variation in $\bar{D}/K$ (or more generally, in $D_i/\bar{D}$, where the subscript $i$ refers to the $i$th grain size in a distribution of mean size. P. D. Komar & Z. Li combined Equations (13) and (15) into the form

$$\frac{\theta_{ci}}{\theta_{cr}} = \left(\frac{D_i}{\bar{D}}\right)^{-h}, \tag{16}$$

where $\theta_{cr}$ is the reference Shields parameter at $D_i/\bar{D} = 1$. The exponent $h$ and $\theta_{cr}$ are functions of $R_*$, but most workers assume that they are constants for fully rough boundaries. The coefficients have been evaluated in flume experiments by Day (1980) using quartz sand-gravel mixtures and in field measurements in quartz gravel-bed streams by Parker et al (1982) and Andrews (1983) (Table 2). Day found that $\theta_{cr}$ equaled the conventional Shields parameter, whereas Parker et al and Andrews obtained a constant value almost double that. The disparity among $h$ values is also important and deserves further study. Day's mixtures were finer grained and possibly yielded different coefficients in Equation (15) as a result of grain size effects or different coefficients in Equation (16) as a result of varying boundary Reynolds numbers. Regardless of these differences among studies, the trends in Equation (16) are clear. Critical Shields parameters of sand-gravel

**Table 2** Representative values of coefficients in Equation (16) for sand-gravel mixtures

| $\theta_{cr}$ | $h$ | References |
|---|---|---|
| $\theta_{c\bar{D}}$ | 0.53 | Day 1980 |
| 0.0876 | 0.982 | Parker et al 1982 |
| 0.0834 | 0.872 | Andrews 1983 |

mixtures are increased over their conventional values for fine sizes and decreased for coarse sizes; thus all sizes become more nearly equally mobile.

There must be some asymptotic limit to the decrease in $\theta_{ci}$ with increasing $D/K$ or $D_i/\bar{D}$. Fenton & Abbott's (1977) data suggest a value of $\theta_{ci}$ equal to 0.01; Ramette & Heuzel (1962) and Andrews (1983) suggest 0.02. Also, with ever-increasing $D/K$ values, the flume experiments by Raudkivi & Ettema (1982) demonstrate that increasingly more vigorous wakes are shed off the larger particles. Scour holes are created on their downflow sides and the particles become embedded, which thereby decreases their protrusion and increases their $\theta_{ci}$. In experiments using binary mixtures, $D$-sized particles embed almost immediately if $D/K > 17$.

It is possible to explain entrainment sorting of size-density mixtures qualitatively by combining these ideas in two diagrams, one representing pebbly sand (Figure 7) and the other representing sandy gravel (Figure 8). In each figure the conventional Shields entrainment curve for quartz is presented for comparison with the entrainment curves for mixtures. The latter are calculated by combining Equations (13) and (15) and ignoring the variations in $\beta_1$, $\beta_2$, and $C_d$ with $R_*$. As expected, the critical Shields parameters for mixtures are higher for grains finer than the mean and lower for grains coarser than the mean when compared with the conventional Shields values (Figures 7, 8). The shapes of the two entrainment curves for mixtures are different because the coefficients in Equation (15) depend upon mean grain size, as shown by Z. Li & P. D. Komar (see Refs. Added in Proof). The most easily entrainable size (minimum with respect to the boundary shear stress isograms) in the pebbly sand mixture is about 0.7 times the mean grain size. Flume data due to Day (1980) are in reasonable agreement with this shape. In the sandy gravel mixture, a wide range of sizes ($0.35 < D_i/\bar{D} < 2$) possess the same minimum boundary shear stress. The upper bound of this zone could be as high as 5 depending upon the true minimum $\theta_{ci}$ for coarse fractions of mixtures. These results are consistent with those of Everts (1973), who found in flume experiments using binary mixtures that quartz spheres either less than 1/2 or greater than 3 times the bed size would not stay in transport at a $\theta_{ci}$ just less than that of the bed.

Five different grain behaviors important to entrainment sorting are noted in Figures 7 and 8. Trapping contributes to entrainment sorting if grains are already in motion and a flow's boundary shear stress plots just below one of the $\theta_{ci}$ curves. In this situation, denser particles may be trapped or selected out of the waning bed load in preference to less dense particles because the denser particles test the bed more often and once in place are not reentrained. The efficiency of this process is probably low, although it may explain some of the gold concentrations in thick, massive,

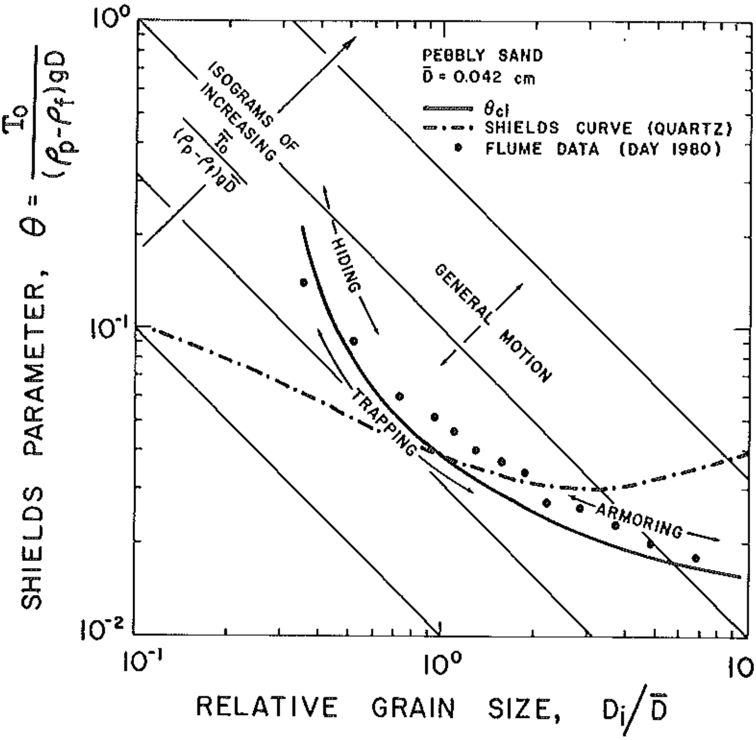


*Figure 7* Threshold conditions for entrainment of grains from a pebbly sand mixture. The ordinate is the Shields parameter $\theta$, in which the grain size is $D$ for the conventional Shields curve and $D_i$ for the $\theta_{ci}$ curve; the abscissa is the relative grain size $D_i/\bar{D}$. Isograms of constant dimensionless boundary shear stress appear as lines of slope equal to $-1$ and possess values equal to $\theta$ at their intersection with the $D_i/\bar{D} = 1$ line. Here $\bar{D}$ equals 0.042 cm, the mean size of the mixture from which Day's (1980) flume data were obtained. The entrainment curve for mixtures is $\theta_{ci} = 0.026 \tan [61.5(D_i/\bar{D})^{-0.3}]$, derived from Equations (13) and (15) in the text. Important grain behaviors in entrainment sorting are (*a*) hiding, in which grains smaller than the mean size require larger boundary shear stresses because of larger reactive angles; (*b*) armoring, in which grains coarser than the mean size experience higher boundary shear stresses because of larger grain weights; and (*c*) trapping, in which grains of size $D_i/\bar{D} \sim 1$ already in motion at a $\theta$ just less than the $\theta_{ci}$ curve are captured in sites on the bed.

crudely horizontally stratified conglomerates of the Ventersdorp Contact placer (Krapez 1985).

Overpassing or winnowing, corresponding to Case 1 of Slingerland (1977), contributes to heavy-mineral enrichment because mode-sized and less dense particles are preferentially entrained at lower boundary shear stresses than are finer- and coarser-sized and denser particles. If the pulp consists of finer heavy and intermediate-sized light minerals—for example,

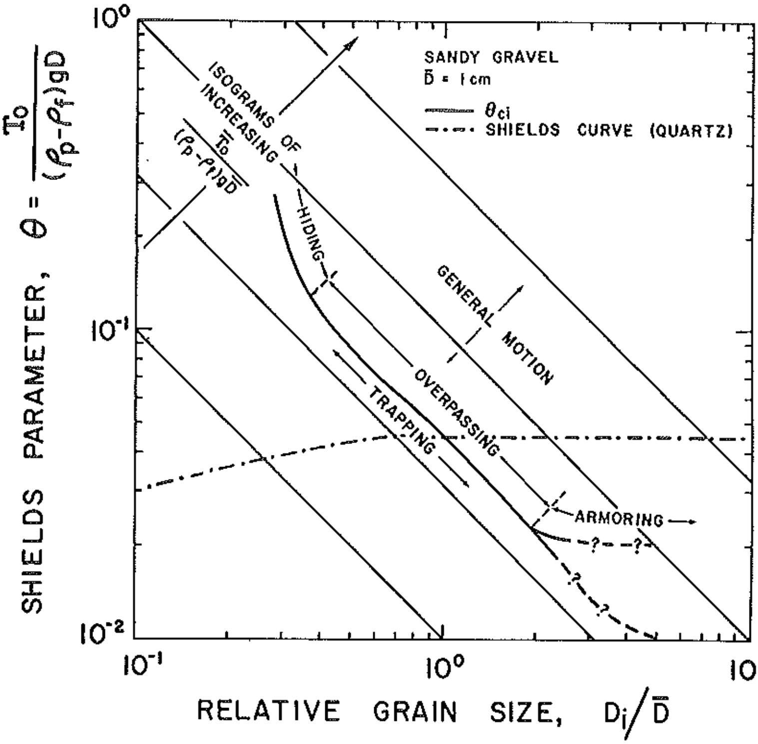


*Figure 8* Threshold conditions for entrainment of grains from a sandy gravel mixture. See Figure 7 for details. Here $\bar{D} = 1$ cm and $\theta_{ci} = 0.0643 \tan [35(D_i/\bar{D})^{-j}]$, where $j = 0.9$ if $D_i/\bar{D} > 1$ and $j = 0.6$ if $D_i/\bar{D} < 1$. In gravel mixtures, grains from one third to twice the mean size possess roughly the same critical boundary shear stress and will overpass, whereas finer and coarser sizes will not.

settling-equivalent size distributions where the light-mineral distribution defines $\bar{D}$—then the bed should become enriched in heavy minerals, and thus the settling velocity ratios of the grains, $w_h/w_l$, should become greater than 1, a condition often observed at the top of beach swash zones (Slingerland 1977, Hand 1967, McIntyre 1959, Stapor 1973, Komar & Wang 1984) and suggested by the data of Stavrakis (1980; his "mixed lamina") from an ancient fluvial deposit. Brady & Jobson (1973) invoke a similar mechanism for heavy-mineral segregation on dune crests.

Armoring (corresponding to Case 3 of Slingerland 1977) is important in entrainment sorting because the finer grains in a mixture can be winnowed away, leaving the coarse tail. As mentioned previously, large clasts of size $D$ embed themselves rapidly in a bed of size $K$ when $D/K \geq 17$ or, for a mixture of sizes, when the geometric standard deviation of the mixture is greater than $\simeq 1.3$ (Little & Mayer 1976). The amount of enrichment occurring by this process depends upon the sizes in the pulp—specifically, the spread between the means of the heavy- and light-mineral distributions. If the mean size of the heavy minerals is similar to $\bar{D}$, the armor may become impoverished and the sediment load enriched because turbulent wakes around large immobile grains may entrain and suspend all the finer (and consequently also denser) particles. The increased $\theta_{ci}$ resulting from the increased density of the particles is less than the decreased $\theta_{ci}$ resulting from their lesser $D_i/\bar{D}$. As the mean size of the particles increases, the increased $\theta_{ci}$ due to increased density may come to predominate, and the armor may become enriched. Concentrations immediately underneath armor layers on top of gravel bars and pebble stringers in ancient deposits [described by Toh (1978), Smith & Minter (1980), and Krapez (1985)] can be explained in this manner.

In hiding, both smaller sizes and greater densities act to increase a grain's resistance to motion. Fine heavy and light grains come to rest among fixed-roughness elements in a greater ratio than in the transported load. Hiding can explain the increased concentrations of heavy minerals, especially gold, in open framework gravels (Smith & Minter 1980).

General motion does not contribute to entrainment sorting per se, because all particles are moving. The rates of motion are unequal, however, and lead to transport sorting by size and density. This process is considered next.

## *Differential Transport of Grains*

Transport sorting results when one size or density fraction of a sediment mix is transported at a different rate from another and so may come to rest at a different location. It involves not only differences in grain transport velocities but also entrainment sorting, since bed-load particles tend to frequently "stop and go" while moving downstream. (Each "stop" requires reentrainment.) Size sorting of quartz-density grains due to differential bed-load transport has been well studied by Gilbert (1914), Einstein & Chien (1953), Egiazaroff (1965), Meland & Norrman (1966), Gessler (1971), Bridge (1981), and Parker et al (1982). Few researchers, however, have studied transport sorting as a function of density, although this is an important mechanism for generating heavy-mineral enrichments (Meland & Norrman 1969).

As with entrainment, key variables in transport sorting are the means and variances of the light- and heavy-mineral size distributions, the bed roughness, and some measure of the fluid force. The phenomenon has been treated theoretically by Slingerland (1984) by using H. A. Einstein's bed-load function, which allows for grain shielding effects. Slingerland calculated transport rates for the different size fractions in a 90% quartz, 10% magnetite sand mix under different combinations of bed roughness and shear velocity. The starting size distributions of the quartz and magnetite were constructed to yield nearly equal settling velocity distributions. Flow strength was calculated by using Equation (2) for various hypothetical depths and slopes. He found that for a given shear velocity, transport rates for all sizes and both densities decrease with increasing roughness, and that the relative proportion of magnetite in the moving bed load increases with increasing $U_*$ for a given roughness and decreases with increasing roughness for a given $U_*$ (Figure 9). These results are corroborated by experimental data (Meland & Norrman 1966, Steidtmann 1982). Meland & Norrman (1966) conducted flume experiments in which transport velocities of glass spheres ($\rho_p \sim 2.54$) were measured over fixed beds of ideally packed, perfectly sorted spheres. Diameters of both the transported and fixed spheres were varied from 0.21 to 0.78 mm to test the effects of relative roughness on transport rates. Results showed that for any particle size and roughness within the experimental range, transport velocity increased with increasing shear velocity, but that for any roughness size, larger grains moved faster than smaller ones (Figure 10). For a given shear velocity, the fastest transport rates were shown by the largest grains (0.78 mm) moving

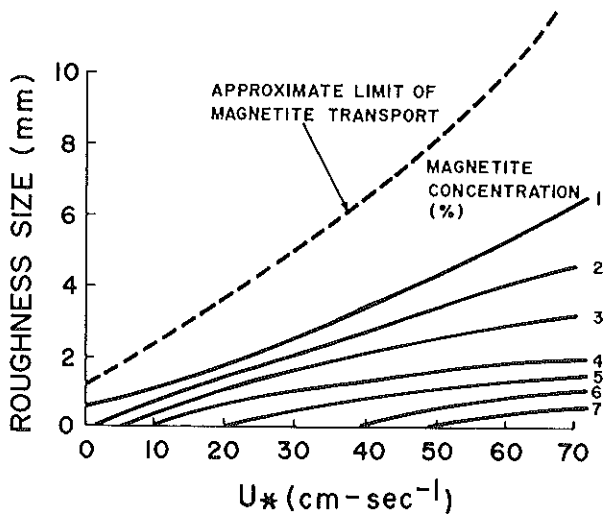


*Figure 9* Effects of bed roughness and shear velocity $U_*$ on the magnetite concentration in the sediment load of a stream as predicted by the Einstein bed-load function (after Slingerland 1984). The bed is composed of 10% magnetite and 90% quartz, with mean sizes of 0.02 cm for the magnetite and 0.042 cm for the total distribution. Maximum bed enrichment of magnetite occurs for higher roughnesses and lower $U_*$.

over the smallest roughness (0.21 mm). As discussed in the section on entrainment sorting, this is a commonly observed phenomenon that is caused by the larger particles projecting farther upward into the velocity profile and rolling more easily over the relatively smooth (small roughness) surface. Conversely, the lowest transport rates were shown by small particles (0.21 mm) moving over the roughest bed (0.78 mm), a result of shielding and higher reactive angles. Because Meland & Norrman did not vary grain density, the significance of this result to heavy-mineral sorting is indirect (e.g. for cases where initial heavy and light grain size distributions are not alike).

In somewhat similar experiments using glass spheres and fixed bed roughness, Steidtmann (1982) investigated the effects of density on particle transport rates (Figure 11). He observed that grains less than 0.2 mm ($D/K$

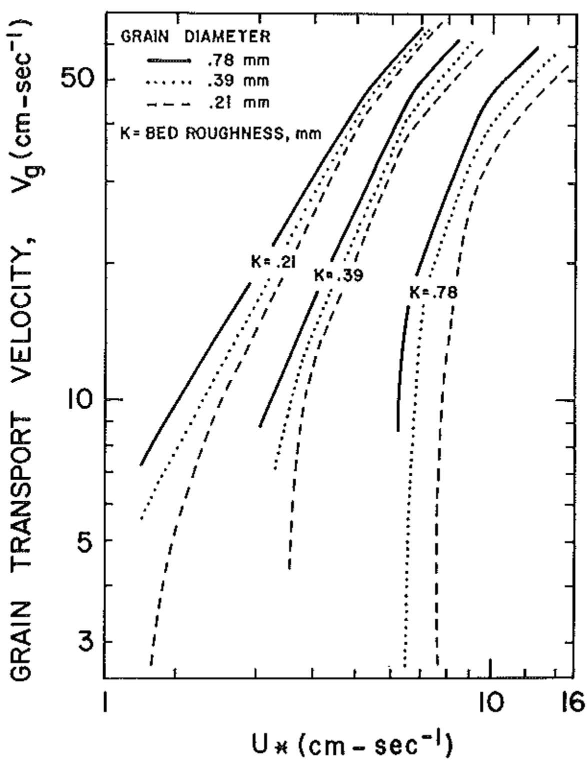


*Figure 10* Relationships between grain transport velocity and shear velocity ($U_*$) for equal-density spheres over fixed beds of perfectly sorted spheres. For a given shear velocity and roughness size, larger grains are transported faster than are smaller grains. Also, note that all particles move slower with increased roughness size (after Meland & Norrman 1966).

< 0.6) are not transported at the shear velocities used in his experiments, a finding that corroborates the results of Everts (1973). This is a consequence of the high reactive angle in Equation (13) and therefore is dependent more on size than density. Grains of larger relative size have transport velocities $V_g$ that increase with both increasing $D/K$ and increasing shear velocity. Transport velocities of the light and heavy minerals are nearly equal for smaller $D/K$, diverge for intermediate $D/K$, and probably converge again at the largest $D/K$. This is because $V_g$ is a measure of a grain's velocity while moving (proportional to its average elevation off the bed), the number of times it tests the bed for a stable resting place, its duration there, and its entrainability. All four factors depend upon a grain's free settling velocity. At the values of $U_*/w_\infty$ used in Steidtmann's study, small heavy and light grains probably traveled high enough in the flow to seldom test the bed and therefore experienced almost continuous motion. Intermediate-size heavy grains, because of their smaller $U_*/w_\infty$ values, probably traveled in closer contact to the bed in the zone of lower fluid velocity (Figure 1), tested the

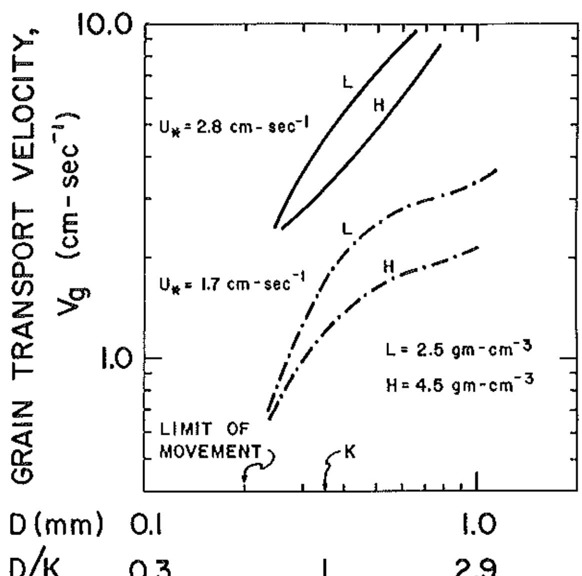


*Figure 11* Generalized relationships among grain transport velocity, grain size $D$, shear velocity $U_*$, and roughness size $K$ for light (L) and heavy (H) spheres moving over a bed of fixed roughness. Curves are drawn through the flume data of Steidtmann (1982). For particles significantly smaller than the bed roughness size, there is little difference in transport velocity between equal-sized light and heavy grains because both types travel in intermittent suspension. For particles near the roughness size, light grains move significantly faster than heavy grains of the same size because heavy grains must be repeatedly reentrained.

bed more times for stable sites, and possessed higher $\theta_{ci}$ compared with equal-sized light grains. One would expect that heavy and light grains with very large $D/K$ would again possess similar transport velocities, because once these grains are accelerated, the forces needed to keep them in motion are equivalent. Steidtmann extended his experiments to include bulk transport rates in which the moving sediment defined its own roughness (i.e. mobile instead of fixed bed). He obtained qualitatively similar results for a plane-bed condition—heavy grains were found in decreased concentrations relative to light grains in downstream deposits. For a rippled bed, however, no systematic downstream differences in transport rates were observed, presumably because of sorting and transport complexities associated with irregular bedforms (Brady & Jobson 1973), but possibly also because of the restricted flume length and limited sampling intervals used by Steidtmann. Certain enigmatic results of an earlier flume study by Minter & Toens (1970) are now more understandable in light of the above. Using a sediment composed of quartz and magnetite grains, with median diameters of 0.57 and 0.083 mm, respectively, Minter & Toens observed that when the sediment was transported over a porous gravel bed, the sediment trapped by the gravel contained a lower magnetite concentration than was contained in the moving bed load. Of course this result contradicts intuition. Apparently, the much smaller sizes of the magnetite fraction, although denser, moved over the very rough gravel bed higher in the flow and therefore tested it less than the larger quartz grains.

How does differential transport form placers? An aggrading bed can be enriched or depleted in heavy minerals relative to the bed load, depending upon $U_*$ and the sizes of the heavy- and light-mineral grains relative to the roughness-determining size. Enrichment of the bed is most intense when $\bar{D}_h$ and $\bar{D}_l$ are near the roughness size and $U_*$ is such that the heavy grains travel with more bed contact than the light grains. Defining the $U_*/w_\infty$ values that effect this must await further study. If the heavy and light minerals possess fall-equivalent diameters, the heavy minerals will always lag behind the light minerals and thus will produce enriched deposits in the upflow portions of the transport path.

## *Shearing of Grains*

Theoretical and experimental studies by Bagnold (1954, 1956) showed that when a concentrated flow of cohesionless particles is sheared by gravity or fluid forces, grain interactions create a force perpendicular to the plane of shearing such that the granular mass expands toward the free surface (i.e. away from the bed). Bagnold called this force the "dispersive pressure." He showed that dispersive pressures are greater on larger and denser grains than on smaller or less dense grains in the same horizon of a grain flow; this result suggests that larger or denser grains would migrate upward toward the surface of a nonuniform sediment mix. "Shear sorting" (Inman et al 1966, p. 800) refers to this vertical fractionation of particles caused by dispersive pressures in a moving granular dispersion.

Sallenger (1979), assuming that the magnitudes of the dispersive pressures predicted by Bagnold also hold for sediment populations of mixed sizes and densities, proposed that the diameters of heavy and light grains in the same horizon of a sheared granular mass would be governed by their densities as

$$d_{\mathrm{h}} \approx d_{\mathrm{l}}(\rho_{\mathrm{l}}/\rho_{\mathrm{h}})^{1/2}. \tag{17}$$

Thus, the diameters would be independent of fluid density, which implies that the same relation would hold in air or water. Equality is approximated because several variables in Bagnold's equation were only assumed, but not demonstrated, to be correct for nonuniform sediment. Sallenger showed that dispersion-equivalent sizes could be expected in such deposits as grain flows (e.g. on avalanche faces of dunes) and beach swash zones, and furthermore that both heavy-mineral laminations and inversely graded beds could be formed by this process. Clifton (1969) had earlier invoked a similar interpretation for inversely graded heavy-mineral-rich beach laminae.

## APPLICATIONS

Our discussion of hydraulic sorting mechanisms has been at the sediment grain scale (millimeter to centimeter), whereas economic placers require enrichments on much larger spatial scales. These enrichments come about by many means, but principally by numerous geomorphological dressing mills operating simultaneously over a large area and migrating slowly in space. The following examples show the types of enriched zones left behind.

Small-scale enrichments frequently are associated with asymmetrical bedforms; commonly, heavy-mineral segregations occur on the stoss sides, crests, and slip faces of active ripples and dunes (Brady & Jobson 1973, McQuivey & Keefer 1969) and in foresets and trough surfaces of dune-formed cross-beds (McGowen & Groat 1971, Smith & Minter 1980, Buck 1983). Such enrichments by bedform activity may combine to attain economic grades at regional scales (Theis 1979, Buck 1983). It is possible that all four sorting mechanisms described above interact to produce heavy-mineral segregations in dunes and ripples (Slingerland 1984, Brady & Jobson 1973). Over the upstream-dipping stoss sides, bed shear stress (and $U_*$) is greatest and the intensity of turbulence is lowest. Larger light grains are preferentially transported to the dune crest, leaving the smaller, slower-moving heavy grains behind either by entrainment transport or by shear sorting. Foreset segregation may result from shear sorting of grain avalanches down the slip face of the migrating bedform. Concentrations lining scour surfaces of trough cross-beds probably result from a combination of entrainment and settling sorting as grains brought to this area by avalanching and suspension settling are reworked by high turbulence and backflow caused by flow separation. [For reviews of bedform mechanics, see Middleton & Southard (1977).]

Planar or horizontal stratification may make up major portions of fluvial deposits, and heavy-mineral-rich laminations are frequently observed in such deposits (Ljunggren & Sundborg 1968, Brady & Jobson 1973, Stavrakis 1980, Lucchitta & Suneson 1981, Buck 1983, Cheel 1984). As is the case for ripples and dunes, no single sorting mechanism produces all plane-bed segregations. Slingerland (1980) suggested that as the boundary Reynolds number is increased over a plane sand bed, the dominant sorting process progresses from entrainment to settling to dispersive (shear) sorting. Recent flume observations by Cheel (1984) indicate that transport sorting is important as well. The grain-size data of Stavrakis (1980, his Figure 6) suggest at least two sorting mechanisms for the heavy-mineral laminae he describes.

Much of the past and present placer mining has been aimed at intermediate scales of concentration. Representative examples are given by Crampton (1937), Boggs & Baldwin (1970), Cobb (1973), Komar & Wang (1984), and Smith & Minter (1980). Commonly, economic placer concentrations occur in the interstices of well-packed gravels, which in modern streams often define specific topographic features such as bars, channel junctions, bends, and riffles. Such deposits probably derive from entrainment and transport sorting of fine bed load as it passes over a rough bed, as suggested by Figures 8 and 9.

A type of intermediate-scale segregation in fluvial sandy sediment is described by Smith & Beukes (1983) and is illustrated in Figure 12. Narrow channelways confined by a bank on one side and a bankward-moving bar on the other were found to be enriched in heavy minerals when the channel was constricted to the narrowest extent allowed by the converging flows. High concentrations of magnetite and chromite, up to five times the background concentrations, appear to result from transport sorting. Transport-equivalent heavy and light grains supplied to the channelway by the bar are subjected to increased shear stresses and strong turbulence in the combined flow of the channelway; the largest light grains are rolled out quickly, and light grains of nearly equal size to the heavy grains are momentarily suspended by turbulent eddies and then transported away, leaving behind a slower moving mass of heavy grains only slightly smaller than the remaining light grains. The situation described by Smith & Beukes [and earlier by Smith & Minter (1980) in a case study of Witwatersrand gold/uranium concentration in conglomerate] may be only a special case for converging flow features in general. In flume experiments, both Wertz (1949) and Mosley & Schumm (1977) observed magnetite enrichments in the downstream ends of channel scours; in the latter study, the scours were formed by flows converging at channel junctions. The scours are probably similar to the bar-to-bank channelways in that turbulence is high and local shear stresses are at or above threshold values for most of the bed load.

Beach placers are another example of intermediate-scale segregation (Rao 1957, Nelson & Hopkins 1972, Stapor 1973). The dressing mill is the inner surf zone (Slingerland 1977, Komar & Wang 1984), and the sorting process is a combination of (*a*) free settling of heavy and light grains out of landward flows and (*b*) entrainment and transport of the larger light grains by seaward flows. Seasonal erosion-deposition cycles enhance the process by providing repeated deposits of near settling-equivalent sizes that can be beneficiated during erosive phases.

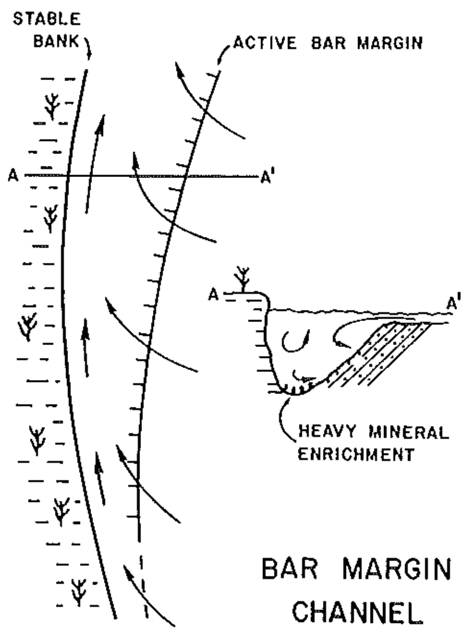


*Figure 12* Sketch showing the generalized geometry and flow patterns formed by an active bar migrating toward a stable bank. Heavy minerals concentrate in the constricted channelway, which gradually widens in the downcurrent direction to accommodate increased flow from over the bar surface. Figure summarized from Smith & Minter (1980) and Smith & Beukes (1983).

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