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Sediment Transport by Wind and Water

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In connection with the 50th anniversary of Aarhus University, 11 September, 1978, brigadier R.A. Bagnold, F.R.S., was conferred the honorary degree of Doctor of Science. The present paper is the manuscript of the lecture R.A. Bagnold was giving at the faculty of science. In the lecture R.A. Bagnold expreses his views on several aspects of sediment transport in air and water. Among the most famous works by R.A Bagnold are the monograph *The Physics of Blown Sands and Desert Dunes,* 1941 (Chapman & Hall, London) and the paper Flow of Cohesionless Grains in Fluids, 1956 *(Royal Soc.London, Phil.trans.A.249,* 235-297). Many of the thoughts presented in the present paper have their roots in these two works.

The effects of the transport of granular solids by both wind and water flow have long been of interest to Earth scientists, in the guise of Sedimentary Geology. The erosional and depositional effects of transport by water flow have since time immemorial been the concern of river and canal engineers. But the dynamic mechanisms involved in the transport processes have until very recently been among the least understood of natural phenomena. Indeed it is doubtful whether modern textbooks of river and canal engineering convey any clearer understanding of the natural processes involved than was probably possessed by the great engineer Pharaohs of the 12th Dynasty 4000 years ago, one at least of whose vast canals appears to have remained self-clearing for 1500 years.

One can attribute some of this backwardness to the fact that the training of neither the earth scientist nor the hydraulic engineer has included any general grounding in the basic principles of physics. Further, the hydraulic engineer, who alone has had the facilities for experiment, is expected not to do scientific research into natural processes, but to exploit their effects for Man's immediate benefit. His traditional approach to the prediction of the rate of water flow is kinematic. Water being of constant density, the elimination of mass and density is a legitimate kind of shorthand, but clearly any kinematic approach to processes of sediment transport must be self-defeating, because it disregards the essential attribute of a sediment – that it is heavier than the fluid transporting it.

From the more aloof viewpoint of the physicist we see the following indisputable facts. Firstly, as I have said, the granular solids are heavier than the fluid, whether air or water, yet when transported by the fluid they are somehow maintained statistically dispersed upwards, to some degree, above the surface of the gravity bed. So some kind of upward force must be exerted between the solids and the bed. Secondly, in all practical conditions the fluid flow is turbulent; that is, in contains random internal eddy currents moving in all directions. Thirdly, a duality of behaviour has long been recognised. Sand grains behave very differently from finer dust particles in a wind. Similarly, sand and all larger grains behave very differently from fine silt or clay particles in a river.

Clouds of fine dust can be raised from dry disturbed ground, carried to great heights by a strong turbulent wind and transported great distances before being finally scattered far and wide over land and sea. Similarly, fine particles tend to be dispersed throughout the body of a river flow. There can be little doubt, and it is generally accepted, that these fine particles are maintained in upward dispersion, or 'suspended' by the internal movements of the fluid turbulence.

In contrast, even during a severe sand storm over a desert dunefield the dense cloud of flying grains rises little more than 1 to 1.5 metres above the dune surface, leaving the heads and shoulders of men projecting against a clear blue sky. The random plumes characteristic of turbulent suspension are conspicuously absent. Indeed I have noticed on such rather unpleasant occasions that the natural gustiness of the turbulent wind is markedly reduced. The sand cloud seems to damp out the violence of the turbulent eddies. Similarly, in many mountain streams, through which vision is but slightly obscured by suspended particles, one can see sand and larger grains being transported as a cloud of 'bedload' restricted to a height of a few grain diameters only above the bed. Again, contrary to the behaviour of dust in the air and of silt or mud in rivers, sand and larger grains tend to accumulate into discreet raised dunes or river bars.

Thus another, different, dynamic mechanism must also operate independently of suspension by fluid turbulence. G. K. Gilbert, around 1910, noticed in a model water flume that grains moved close over a gravity bed in successions of regular jumps. He called this motion 'Saltation'. No attempt was made to explain the dynamics of this rather curious phenomenon. Later, in the 1930's, I showed that wind-blown dune sand grains moved entirely in saltation, fluid turbulence having little or no lifting effect.

Saltation is a very general phenomenon. A simple example is that of a rubber ball driven bouncing along by a strong wind. The upward lift, or in the case of a swarm of like balls, the statistical upward dispersion, come from successive instantaneous reactive impulses imparted at each contact with the bed. Elastic rebound is not necessary. It may or may not occur. When a stone is made to skim over a water surface it may make several saltations before sinking. The upward impulses are then purely fluid-dynamic.

The French term for saltation is 'charriage'. But they appear to have overlooked the implications of this strangely apt word. When an un-sprung cart (char) is pulled rapidly enough across a ploughed field it must saltate from ridge to ridge. But the average weight which the ground supports remains precisely the same whether the cart saltates or not. i.e., whether the load is continuous or intermittent.

The basic mechanism of saltation is very simple in terms of grain momentum. A descending grain has a momentum component mU_1 in the forward direction. Whether the impact is elastic or inelastic its subsequent forward momentum is

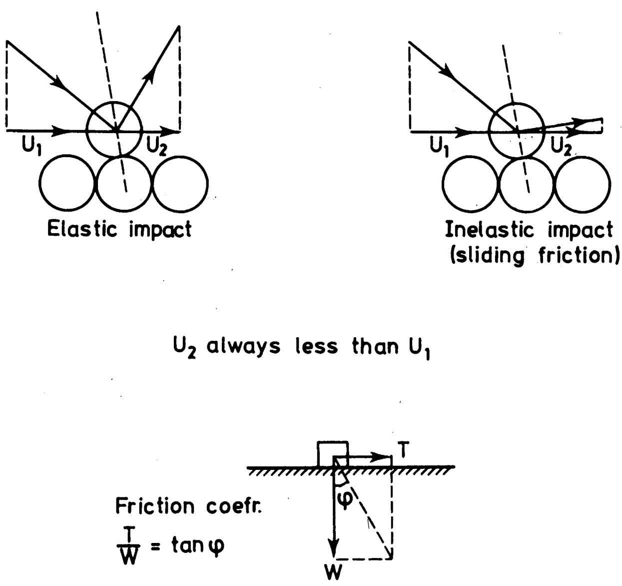

Fig. 1.

smaller at mU_2 some having been transferred to the bed. So in static fluid conditions successive trajectories decrease in magnitude till motion ceases. (Throwing a ball in still air). To maintain the saltation a mean tangential fluid thrust is necessary, to accelerate the grain during each salation from velocity U_2 to velocity U_1.

Thus we have a fluid thrust of a steady time-mean value T in the tangential direction, and a perpendicular downward thrust on the bed of a time-mean value m equal to the grain's immersed (excess) weight. The statistical ratio T/m is clearly in the nature of a dynamic friction coefficient of the same kind as that between two solids in continuous contact. The value of the latter coefficient is usually expressed in terms of the tangent of the limiting friction angle φ, so I have called the dynamic friction coefficient for intermittent saltating contact $\tan \alpha$. An experiment I did some 20 years ago showed $\tan \alpha$ to be of the same order as the limiting static coefficient for $\tan \varphi$ for mineral sands. More recently Francis from analyses of high speed multi-image photography of grains saltating in water, has confirmed this, putting the value of $\tan \alpha$ at approximately 0.6.

The importance of this is that it gives a simple relation between the immersed weight stress of the solids transported as bedload and the shear stress needed to be applied by the fluid flow to keep the bedload moving. For instance, if the fluid shear stress applied to the saltating grains is, say, 0.6 kg per square metre, the immersed weight stress of the bedload in transit will be 1 kg per square metre approx. This removes much of the mystery created by the traditionalists who by their insistence on the kinematic approach ignore the effects of frictional resistance. The present dynamic approach on the other hand brings the mechanism of bedload transport into the realm of everyday experience.

We can regard the fluid flow, whether air or water, as a transporting machine which does mechanical work in pushing the bedload along against a frictional resistance. We have at once a simple relation between the rate at which bedload is transported, in terms of immersed weight passing per second per unit width, and the rate at which fluid kinetic energy is dissipated in the process. If the transport rate of the bedload is i_b kg per second per metre width then by definition

$$i_b = W'_b \bar{U}_b$$

where W'_b is the immersed bedload weight per square metre and $\bar{U}_b$ is its mean travel velocity. (Note: In dealing with two-phase – fluid-solid – dynamics it is necessary to use separate symbols for the velocities of fluid and solid at the same point because relative velocities are involved. It is convenient to use the capital letters U and V for the velocities of the solids, reserving small u and v for the corresponding fluid velocities).

Since $W'_b = T/\tan \alpha$, where T is the fluid stress exerted on the saltation, and the work rate is $T\bar{U}_b$, we have

$$i_b = \frac{T\,\bar{U}_b}{\tan\alpha} \quad \text{or work rate} = i_b \tan\alpha$$

Now for any continuously working machine the conservation of energy enables us to write

Work rate = input power × efficiency

The directly applied input power or rate of energy supply is the product of the fluid stress T exerted on the saltation multiplied by the fluid velocity u_y within the saltation zone, measured at a height y analogous to that of a 'centre of thrust'. Thus

$$\text{directly applied fluid power} = T\,u_y$$

So our general machine equation becomes

work rate = input power × efficiency

$$i_b \tan\alpha = T\,u_y \times E = T\,\bar{U}_b$$

Now a fluid thrust on an immersed body involves a mean relative velocity u_r between fluid and solid, so we can substitute $(u_y\text{-}u_r)$ for the mean travel velocity $\bar{U}_b$ whence $T(u_y\text{-}u_r) = Tu_y \times E$ and Efficiency $E = (u_y\text{-}u_r)/u_y$

This notionally introduces the threshold condition at which bed grains cease to be moved by the flow. The efficiency falls to zero when u_y is equal to or less than u_r.

We can go one step further in our direct inferences from general dynamic concepts. The relative or 'slip' velocity u_r must, it seems, always approximate to the terminal velocity V_g of fall of the grains through the water. So we can substitute this easily measurable quantity for the un-measurable quantity u_r. The argument is surprisingly simple once one recognises the existence of the friction coefficient $\tan\alpha$. The equilibrium condition for the constant fall velocity is

upward fluid force = immersed weight W'_b

$$\frac{\rho C \pi D^2}{4} V_g^2 = \frac{\pi D^3}{6}(\sigma-\rho)g$$

$$V_g = \sqrt{\frac{2}{3}\frac{(\sigma-\rho)}{C\rho}\,gD}$$

σ and ρ are the densities of the grains and water, respectively.

Since the tangential fluid stress T is equal to the weight stress $W'_b \times \tan\alpha$, the corresponding tangential relative velocity u_r is equal to $V_g\sqrt{\tan\alpha}$, whence

$$u_r \equiv V_g \sqrt{\tan\alpha} \approx 0.8\, V_g$$

whence $$i_b \approx \frac{T}{\tan\alpha}\,(u_y - V_g)$$

If therefore we could assume that the whole of the measurable fluid shear stress τ is exerted on the saltating grains and none on the stationary bed below, so that $T = \tau$, the prediction of the bedload transport rate i_b would need only one unknown quantity to be evaluated, namely the fluid velocity u_y within the saltation zone. Unfortunately our understanding of 2-phase (liquid-solid) flow in the presence of a transverse gravity stress acting differentially on the solid phase is so meagre that the value of u_y is not only quite unpredictable but also very difficult to measure in practice. So it is necessary from this point onwards to resort to empiricism based on experimental results, and then see whether the empiricisms are mutually consistent and physically plausible.
In particular we have to express the applied fluid power arbitrarily in some easily measurable way. The whole fluid shear stress τ can be readily measured or inferred; and τ can be expressed in terms of the so-called friction velocity u_* *as*

$$\tau \equiv \rho u_*^2$$

And since the power, which we will call omega (ω), has the nature of stress times velocity, omega can be expressed in an arbitrary way as

$$\omega = \rho u_*^3$$

This of course involves the arbitrary assumption that $T = \tau$

Now the conditions in which sand grains saltate in air and water are markedly different, owing to the very great difference in the fluid densities. A sand grain in air is some 2000 times as massive as the fluid surrounding it whereas it is only some 2.6 times as massive as water. A grain saltating in air is so relatively massive that on descent it strikes the bed like a pellet from a shot-gun. On a bed of loose like grains it makes a little crater and splashes a number of bed grains up into the air stream. As these in turn saltate a mechanical chain-reaction is created. The movement of one grain at the threshold power ω_0 of grain movement starts a swarm of other grains in motion, so that the transport rate i_b starts discontinuously at a certain finite value.

Plotting i_b against ω on log-log paper we get a straight line giving the transport rate as remaining proportional to the power ω from the threshold upwards, and apparently constant efficiency. The discontinuous start is very evident. It suggests that owing to the chain reaction the assumption that T approximates to τ may well be a reality.

A sand grain saltating in water, on the other hand, contacts the bed so gently

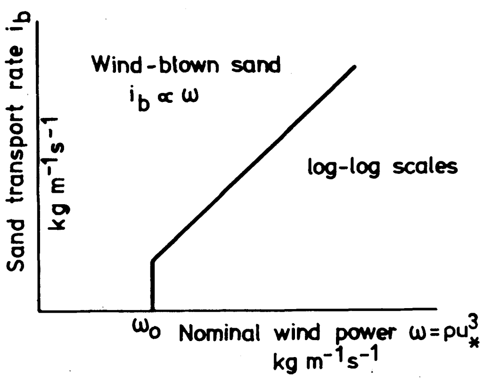

Fig. 2.

that no bed grains are disturbed. The chain reaction is absent. The transport rate i_b therefore increases continuously from zero at the threshold power ω_0. The same log-log plot of i_b against ω for sand in water has in consequence a quite different appearance. It is no longer a straight line. The curvature is however a spurious result of the log-log system of plotting. For we now have two continuous quantities which do not vanish simultaneously. So the transport rate i_b appears to dive towards minus infinity as the power omega approaches its finite threshold value ω_0. This spurious curvature is removed by plotting i_b against a new, excess, power $\omega' = \omega - \omega_0$. The result is a straight line as before, but instead of i_b being proportional to the power it is found to vary as $\omega'^{3/2} = (\omega - \omega_0)^{3/2}$. This is found to be true not only from measurements in a laboratory flume. It appears to hold also, now that we have developed a reliable means of measuring bedload transport rates in natural rivers, for great rivers such as the Snake and others in USA.

A plausible explanation of this different transport rate function is as follows. In

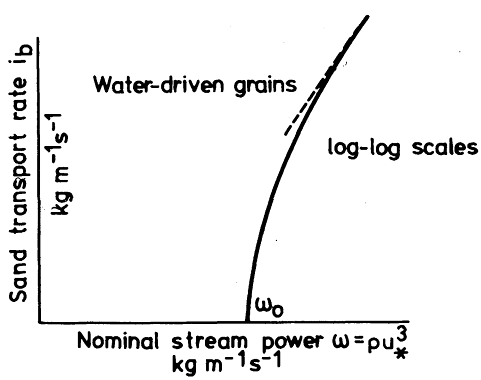

Fig. 3.

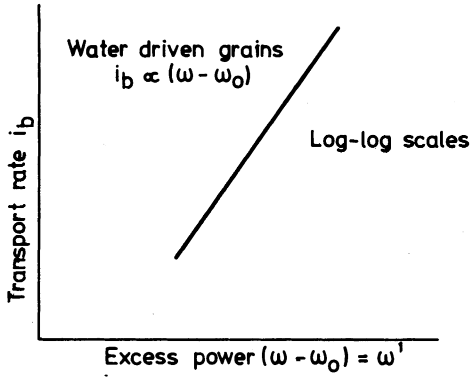

Fig. 4.

the absence of any chain reaction the bedload transport rate i_b increases, as I have said, continuously from zero at the threshold of bed movement. Thus, unlike sand in air, the saltation is negligible at the threshold. So the fluid stress T on the saltation must also be negligible, and the ratio T/τ must increase continuously from zero and with it the transport efficiency. Assuming plausibly that the ratio T/τ increases as some function of ω', the appropriate function should, empirically, be

$$\frac{T}{\tau} \propto \left(\frac{\omega'}{\omega_0}\right)^{\frac{1}{2}}$$

so that

$$i_b \propto \omega' \left(\frac{\omega'}{\omega_0}\right)^{\frac{1}{2}}$$

Now we come to a puzzling phenomenon which I can't yet explain quantitatively; namely the variation of the transport rate with grain size, both in air and in water. Ever since the 1930's when the transport rate of wind-blown sand was measured by Zingg in USA, Chepil in Canada and myself in UK, it has been known that for a given wind power ω the value of i_b increases with increasing grain size. Zingg's measurements, which covered a much wider range of grain size than my own, gave i_b increases as (grain size $D)^{2/3}$ approx. This variation has remained very awkward dimensionally because under the open atmosphere there seems to be no other length with which to form a dimensionless ratio, so the transport rate – wind power relationship has remained dimensionally incomplete.

A flowing liquid, on the other hand, needs to be contained in a pipe or channel. This itself provides the required second length dimension. It has recently become abundantly evident that for a given power $\omega = \rho u_*^3$ the transport rate i_b measured in an open water stream increases, not directly as the grain size $D^{2/3}$, but as the ratio $D/Y^{2/3}$, where Y is the depth of flow. Again, this empirical relation appears

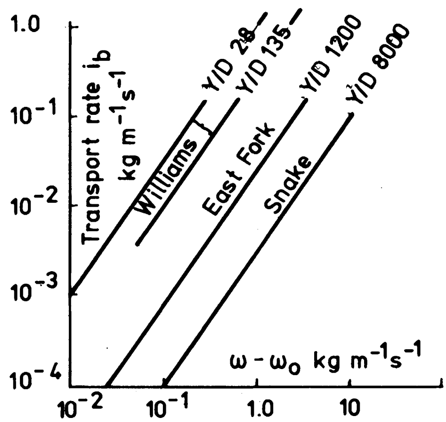

Fig. 5.

to hold over the enormous range of Y/D from G. P. Williams' small scale flume measurements, Y/D = 28 to 135, to W. W. Emmett's measurements on the Snake and other great American rivers, Y/D around 10,000.

I have so far used the arbitrary definition of fluid power as ρu_*^3 in order to compare the simple transport rate function for wind-blown sand with the modified function for water-borne bedload. The real total power of a water stream, per unit bed area, is

$$\omega = \tau\bar{u} = \rho\bar{u}_*^2\,\bar{U}$$

where $\bar{u}$, which can have no definable meaning under the open atmosphere, is the mean flow velocity of the stream, = discharge divided by cross-sectional area.

The total power $\tau\bar{u}$ of a water stream, per unit width, is easy to understand when one considers a unit column of water of height Y descending a gravity slope S at a flow velocity $\bar{u}$. The rate of loss of gravity potential energy is the weight ρgY times the vertical velocity $\bar{u}S$ of descent. The product is equal to $\tau\bar{u}$. Most of this energy is continually being converted first into the random kinetic energy of turbulence and subsequently into heat. Some, however, is converted directly into heat in the process of pushing a bedload along against friction with the bed.

The ratio of the two definitions of flow power ρu_*^3 and $\tau\bar{u}$ is $\bar{u}/u_*$. In simple homogeneous liquid flow $\bar{u}/u_*$ would increase with increasing flow depth roughly as log $12Y/D$. This in itself would be insufficient to account for the general $(Y/D)^{2/3}$ variation. But the data show that the effect of a bedload on the velocity distribution in a water stream is to make the variation of $\bar{u}/u_*$ smaller still. Hence the factual variation

$$i_b \propto \left(\frac{Y}{D}\right)^{-\frac{2}{3}}$$

cannot be explained so simply. Nor, in any case, could such an explanation apply to wind-blown sand, where the flow depth Y of the atmosphere is undefinable.

According to the very wide range of stream data now available, a reasonably reliable empirical expression for the transport rate of un-suspended bedload in water is

$$i_b \equiv \frac{A}{\tan\alpha}(\omega-\omega_0)\left(\frac{\omega-\omega_0}{\omega_0}\right)^{\frac{1}{2}}\left(\frac{Y}{D}\right)^{-\frac{2}{3}}$$

where the power omega is defined as $\tau\bar{u}$ per unit width of the flow. Assuming $\tan\alpha$ to have a value around 0.6 the numerical coefficient A appears as of the order of unity.

The variation of the transport rate with the depth/grain size ratio has an immediate practical implication of some importance. Consider a stream of constant discharge and constant gravity gradient to be in equilibrium, neither aggrading nor degrading. Suppose an industrial development upstream increases the rate of input of sediment into the stream. The stream has the freedom to adapt to a new equilibrium state by increasing its width and thereby decreasing its depth Y. The result is an increase in the transport rate i_b thus allowing the excess sediment input to be carried away.

Again, countless sediment transport measurements have been made in laboratory flumes by hydraulic engineers. Before the introduction of tilting flumes the only way of varying the flow was by varying the flow depth Y. Even when flow gradients could be adjusted, different experimenters used different Y/D ratios. The results are so inconsistent that efforts towards any general correlation were virtually abandoned. It was not until Williams' experiments became known in 1970 that the cause of the inconsistencies became clear. By repeating his measurements in identical conditions at three different constant flow depths he showed conclusively that the transport rate increases progressively as the depth ratio Y/D is reduced, at constant power omega. The available range of flow depth, being only 3-fold, was too narrow to allow of any generalisation. But as Emmett's bedload measurements in American rivers began to come in, a general $(Y/D)^{2/3}$ variation soon became clear to me from comparative plotting. I then went back over existing laboratory data, from Gilbert's massive material of 1914 onwards, applying this Y/D relation to convert the recorded transport rate data to an arbitrary constant Y/D. As I expected, the gross inconsistencies disappeared.

Before leaving the subject of unsuspended transport by saltation, I would point out a peculiarity confined to transport by the open gravity flow of a liquid. The wind which drives sand under the open atmosphere or through a wind tunnel is maintained by an applied pressure gradient in the flow direction. A pressure gradient creates a purely spatial thrust acting on fluid and dispersed solids alike. The gravity component acting on a water stream, on the other hand, acts differentially. It exerts a direct thrust on the excess mass of the solids. Thus the real bedload relation involves replacing $\tan\alpha$ by $(\tan\alpha - \tan\beta)$ where $\tan\beta$ represents the gravity slope of the stream. $\tan\beta$ is usually so small, both in Nature and in laboratory flumes, that it can be neglected in comparison with $\tan\alpha$. But as the slope steepens, in a mountain stream for example, the fluid flow has to do relatively less transporting work. Ultimately, as $\tan\beta$ approaches $\tan\alpha$ in value, that is, as the slope angle approaches the angle of the slip of the bed material, the sediment becomes selv-transporting. The transport rate i_b notionally becomes infinite, an avalanche occurs which ecplains the occurrence of occasional sudden disastrously large sediment movements.

It is but a step from here to the dynamics of mud-flows. The 'fluid' is now a mud whose density ρ' may approach that of the stones or rocks dispersed within it. If the volume concentration of these is C, the critical gravity slope $\tan\beta$ is given by

$$\tan\beta = \frac{C\tan\alpha}{\frac{\rho'}{\sigma-\rho'} + C}$$

The angle β may be reduced to a few degrees only. No such phenomena could occur when gravity is replaced by a pressure gradient.

I have dealt hitherto with the transport of un-suspended bedload moving is saltation in air or water. Note that I have not needed to invoke the mysteries of fluid turbulence to explain anything. It is very evident from the muddy appearance of many streams and rivers and from the huge dust clouds sometimes seen both here and on the planet Mars that fine particles are certainly carried in suspension by the random eddies of turbulence. The conventional textbook treatment of suspension seems to me both misleading and inadequate. The kinematic approach disregards the excess weight of the solids. It treats them as being endowed with the sole property of a downward relative velocity of fall. The suspended solids are in effect assumed to be little fishes of neutral density perpetually swimming downwards at the fall velocity V_g. This approach tacitly assumes (a) the presence of a cloud of suspended solids has no reaction on the fluid flow, (b) the concentration of suspended solids at some basal datum plane is already known, and (c) the random eddy motions are statistically summetrical, the mean upward and downward motions being identical. This conventional theory fails to predict the one useful quantity one wants to know, namely the total load of solids which a given turbulent flow can carry in suspension.

A more realistic dynamic approach must clearly admit both the excess weight W'_s of a suspended load and the necessity therefore of a mean upwardly directeed stress to maintain the suspension. I reasoned that such a stress could not be exerted by a shear turbulence whose eddy motions were symmetrical, and that it was necessary to assume to some degree asymmetry whereby, statistically, relatively small volumes of fluid are moving upwards away from the bed at relatively large velocities $+v'$ and are being replaced by larger volumes moving towards the bed at relatively small velocities - v? Continuity demands that the total fluid momentum component $m\overline{v}$' normal to the bed shall remain zero, so that $m_{up} \cdot v'_{up} = m_{down} \cdot v'_{down}$. But the upward momentum flux $m_{up} \cdot v'2_{up}$ now exceeds the downward flux. The difference constitutes a mean upward stress f capable of exertion on any immersed body.

Analogously to the friction velocity u_* defined as $\sqrt{\tau/\rho}$ we can notionally define the upward propagation velocity of the stress f as $\sqrt{f/\rho}$. Thus the shear turbulence would be capable of doing lifting work at the rate $f\sqrt{f/\rho}$. The work rate required to maintain a load W'_s in suspension is $W'_s \cdot V_g$. As a first approximation we may assume the mean travel velocity $\overline{U}_s$ of the load W'_s to be equal to the mean flow velocity $\overline{u}$. Thus the transport rate i_s of a suspended load will be

$$i_s = W'_s \overline{u} = f\sqrt{\frac{f}{\rho}}\ \frac{\overline{u}}{V_g}$$

where the measurable ratio $V_g/\overline{u}$ for a suspended load has a meaning analogous to the friction coefficient $\tan\alpha$ for the bedload. Notice that since the fall velocity V_g of small particles decreases as the square of their diameter the potential transport is increases very rapidly as the size of the particles is reduced. A fifty percent concentration of very small dispersed clay particles has been recorded in the Rio Puerco river in USA.

Measurements by Laufer and others of the root-mean-square value $\sqrt{v'^2}$ of the eddy velocity component v' indicate that v'_{max} near the shear boundary approximates to the friction velocity u_*. By making a reasonable assumption as to the probable degree of asymmetry, I estimated that the value of i_s should be around

$$i_s \approx \omega(0.01\ \frac{\overline{u}}{V_g})$$

where V_g for heterogeneous particles of a measured size distribution is given by $\sum \rho V_g / \sum \rho$. ρ denotes percentage by weight. Using this relation I plotted the discrepancy ratio predicted value/measured value for 146 existing suspended transport measurements in eight rivers in USA. There was, as I expected, a very wide scatter, but, perhaps by luck, the geometrical mean discrepancy amounted to 75%.

Such is the inertia of conventional thought that the new theory of suspension attracted very little interest. The reason is, I think, that since eddy velocities of turbulence have been measured in terms of local energy change the results have been root-mean-square values only. The symmetry or otherwise of the motions have therefore been undetectable. So it has been taken for granted, without any supporting evidence, that the eddy motions of shear turbulence must be symmetrical. Recently however experiments using the new hydrogen-bubble

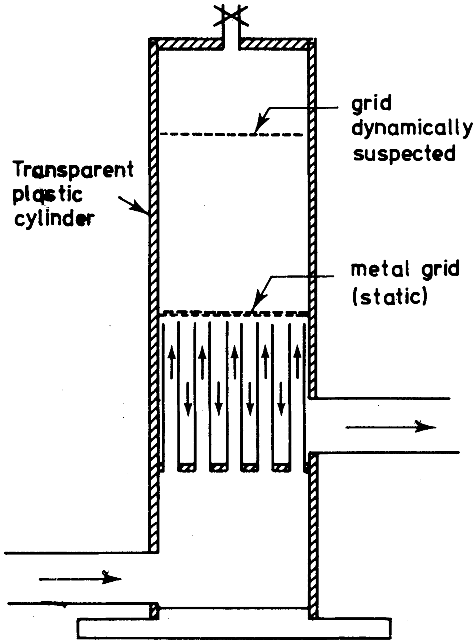

Fig. 6.

technique have disclosed the existence of fast jets of fluid moving upwards away from the shear boundary, strongly suggesting an inherent asymmetry.

The effect of an asymmetrical turbulence can be demonstrated in a rather spectacular way in a simple bit of apparatus I put together last year. The whole system is filled with water. A metal grid lies at rest at the bottom of the upper chamber. The crosssectional area of the vertical tubes below is smaller than that of the space between them. On starting a through-flow of water the metal grid is seen to rise upwards and to remain suspended, seemingly without support, as if levitated.

A statistically upward thrust of this dynamic kind, if it exists in reality, must inevitably be accompanied by an equal and opposite downward thrust exerted on the bed. Such a thrust has in fact been detected as an excess static pressure within the pore-space between the bed grains. Moreover recent measurements have shown the magnitude of this excess pressure to be of just the right order.

Time has forced me to omit many important aspects of sediment transport. I might mention the dependence of the time rate of erosion and deposition on the distance rate di/dx of change of transport rate along the flow. Also the variation of the transport rate of wind-blown sand with the size distribution of the surface grains and hence the reason why sand collects preferentially on a surface of like sand to form a dune. Again, I have omitted any discussion of the important reaction of sediment transport on the structure of the impelling fluid flow. The saltation of wind-blown sand grains causes very marked changes in the velocity pattern of the wind. Similar changes are to be expected in water but accurate measurements are so far lacking.

I have tried to outline the bare bones of the subject as I see them. There is still a very great deal of detail to be filled in by careful experiment designed scientifically with the object of discovering the natural truth rather than merely to further the technological exploitation or the phenomena of Man's immediate benefit.

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