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WOODS HOLE OCEANOGRAPHIC INSTITUTION / MASSACHUSETTS INSTITUTE OF TECHNOLOGY

JOINT PROGRAM IN OCEANOGRAPHY AND OCEAN ENGINEERING

DOCTORAL DISSERTATION

FLOW AND SKIN FRICTION OVER NATURAL ROUGH BEDS

BY

CHRISTOPHER PAOLA

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WOODS HOLE OCEANOGRAPHIC INSTITUTION
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DOCTORAL DISSERTATION

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FLOW AND SKIN FRICTION OVER NATURAL ROUGH BEDS

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Submitted to the Department of Earth and Planetary Sciences, Massachusetts Institute of Technology, and to the Department of Geology and Geophysics, Woods Hole Oceanographic Institution, on February 16, 1983, in partial fulfillment of the requirements for the Degree of

DOCTOR OF SCIENCE

at the

MASSACHUSETTS INSTITUTE OF TECHNOLOGY

and the

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February, 1983

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Thesis Supervisor

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Thesis Supervisor

Accepted by: _____
Chairman, Joint Committee for Marine
Geology and Geophysics

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ABSTRACT

When a boundary layer develops over a bed that is hydrodynamically rough at a length scale or scales larger than the grain size (a macrorough bed), as is usually the case where bed forms are present, it is necessary to distinguish among total boundary shear stress and its components, form drag and spatially averaged skin friction. It is known that the mean-velocity field reflects the composite boundary shear stress. Above about one roughness height above the tops of the roughness elements, the velocity does not vary horizontally. Its vertical profile is semilogarithmic and scales with the total friction velocity $u*_t$ and total roughness length z_{ot}. This region is here called the integrated logarithmic layer (ILL). Below the ILL the velocity varies horizontally in response to the irregular boundary; this region is called the surface layer.

In the first of two sets of experiments reported here, skin-friction measurements were made with an array of flush-mounted hot films at four points on the stoss slope of one of a field of two-dimensional immobile current ripples. Total boundary shear stress was also measured, as were mean-velocity profiles in the ILL and the surface layer. The ILL behaves as described above. Although surface-layer velocity profiles are semilogarithmic, their semilogarithmic slope is not proportional to the local skin-friction velocity, so they do not locally obey the law of the wall. Rather, the velocity field can be decomposed into a spatially averaged rotational component and a local inviscid perturbation. The measured skin-friction field is consistent

with a simple model for sediment transport over the bed forms except near reattachment, where the fluctuating skin friction is important. The data are also consistent with the drag-partition theories of Einstein and Barbarossa (1952) and Engelund (1966). Normalized skin-friction spectra do not vary with streamwise position but do vary with Reynolds number; skin-friction probability density functions show significant increases in skewness and kurtosis near reattachment but do not vary strongly with Reynolds number.

In the second set of experiments the skin-friction vector field was measured around isolated hemispheres, with model sedimentary tails one and four obstacle heights long and without tails. The measured skin-friction fields are not consistent with deposition along the obstacle-flow centerline downstream of reattachment, which occurs about two obstacle heights downstream of the trailing edge of the hemisphere. This applies for local bed-load erosion and deposition and for general deflation of the bed, and is not substantially altered by the presence of either tail. Measurements were also made of skin friction, total boundary shear stress and ILL velocity profiles over h,B-rough arrays of hemispheres with and without tails four roughness heights long, at two areal densities. The skin-friction field in the denser array is significantly distorted from that around an isolated element. The measured skin friction in both arrays is significantly greater than that given by a drag-partition formula proposed by Wooding et al. (1973). The roughness length z_{ot} for both densities is not changed by addition of the tails.

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Dr. Giselher Gust

Title: Associate Professor of Marine Science, University of South Florida

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Contents

List of Figures:

List of Tables

1. Introduction.

1.1. General. Much of physical sedimentology involves consideration of the motion of fluid and sediment over boundaries irregular on one or more scales. Bed forms are among the commonest sources of irregularity, and they are of particular importance geologically because in many cases they produce a sedimentary record that is distinctive and potentially rich in paleoenvironmental information. In the analysis of modern environments, bed forms play an important role as natural current meters; the information they provide is potentially most valuable in places such as the deep ocean where direct measurement of currents is difficult and expensive. Interpreting both the ancient and the modern bed-form records amounts to deducing flow properties from observations of bed-form characteristics. This may be viewed as an inverse problem in which the forward problem is to determine the properties of bed forms developed under a given imposed flow field. This forward problem in itself is extremely complex: the simplest elements to which it can be reduced form a coupled, turbulent system of flow and sediment transport under locally nonuniform conditions. In the face of this it is not surprising that the most productive approach from a sedimentological viewpoint has been direct and empirical (Southard, 1971; Dalrymple et al., 1978; Harms et al., 1982).

As in any branch of science, however, such empirical information must always be supplemented by analysis aimed at providing an understanding of why the results have the form they do. Apart from the aesthetic pleasure such understanding provides, without it we cannot assess the stability of empirical results to changes in conditions from those under which they were obtained. Spatial and temporal variations in scale and magnitude of flow and in sediment properties all affect bed forms in nature; to evaluate the effects of all of these empirically we would have to generate an enormous catalog of data. A more effective approach is one that combines theory with well-placed critical experiments.

So much for putting the general problem of bed-form dynamics into geological perspective; it still needs to be reduced substantially to bring it within range of the available means of attack. I mentioned above that any bed-form theory must involve consideration of both flow and sediment transport over irregular (henceforth "rough") boundaries. I have chosen to concentrate on the flow, because the flow field must be understood at least near the boundary before there is any hope of calculating the sediment transport. There are, however, complex problems relating to the sediment transport as well. Some of them will be discussed in section 2.53, but the main emphasis in this work will be on the flow and the tangential bottom stresses it sets up.

The nature of the boundary shear stress depends on the form of the boundary, as illustrated by the three cases shown in figure 1.1. In all three, it is assumed that the boundary layer is steady, uniform, and unstratified, and that no sediment transport is occurring. ("Boundary layer" will usually be used in a general sense to mean that part of a wall-bounded shear flow in which shear stresses generated near the wall are dynamically important. Occasionally, though, it will be necessary to distinguish among pipe, open-channel and developing boundary layers.) More detailed discussion of most of the following review can be found in Tennekes and Lumley (1972, ch. 5), Arya (1975), Townsend (1976, ch. 5), and Smith and McLean (1977).

Consider first a turbulent boundary layer developed over a perfectly smooth surface (figure 1.1a). (We will see presently that many real surfaces are "smooth enough".) Although the boundary layer is fully turbulent, as we approach the wall the viscosity must become important to satisfy the no-slip condition. Hence the flow near the wall is viscous-dominated; at the wall the Reynolds stress is extinguished and the instantaneous boundary shear stress T_o is given by $T_o=\rho\nu(\partial U/\partial y)_o$, where ρ and ν are the fluid density and kinematic viscosity respectively, and U is the instantaneous velocity at height y above the bottom (figure 1.2). The subscript zero indicates that both the stress and the vertical derivative are to be evaluated at the bed.

In a turbulent flow T_o and U can be divided into time-averaged and fluctuating parts. Nearly all of the flows to be considered in this work will be turbulent; variables describing them will be distinguished as follows. All literal references will be to time-averaged quantities unless stated otherwise. Among symbolic references, lower-case letters or overlining will denote time-averaged quantities, primed lower-case letters will denote temporal flucuations, and upper-case letters will denote total instantaneous quantities. Thus $A=a+a'$ where $a=(1/t_a)\int_0^{t_a} A\,dt)$, and t_a is a time scale long compared with those of the turbulent fluctuations.

Anticipating that τ_o as well as ν may influence the flow field near the smooth wall, we define a kinematic mean shear stress $u*=(\tau_o/\rho)^{1/2}$ having units of velocity and called the friction velocity or shear velocity. Then a natural length scale is $\nu/u*$ and the velocity field near the bed is given by

$$\frac{u}{u*} = f_s\left(\frac{u*y}{\nu}\right) \qquad 1.1$$

Far from the wall, in the outer part of the flow, the turbulence begins to be affected by the finite height of the boundary layer D, so D becomes the length scale. Since the outer scaling is to be viewed as being applied from the surface downward, we refer the velocity to the surface velocity u_s; it should also scale with u*, on the grounds that whatever is the overall driving force on the flow

Figure 1.1. Sketches of smooth, rough, and macrorough flows (left) and corresponding mean-velocity profiles (right). Multivalued profiles are for different streamwise positions.

VDL - Velocity-defect layer

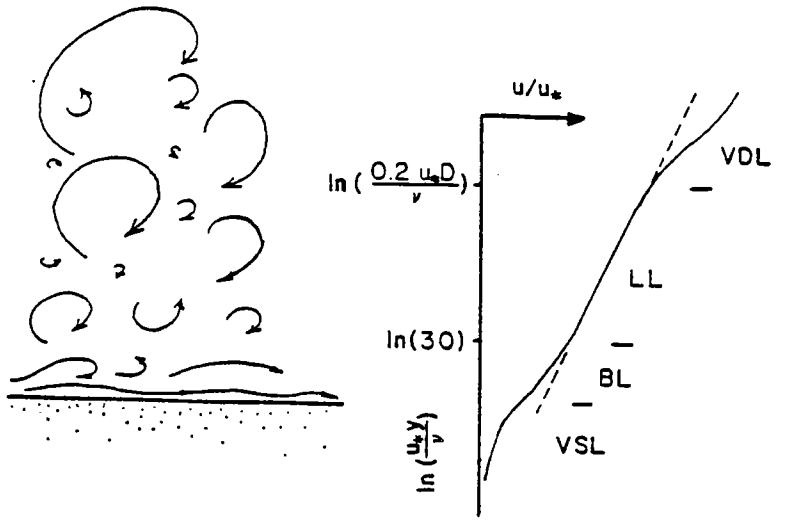
LL - Logarithmic layer

BL - Buffer layer

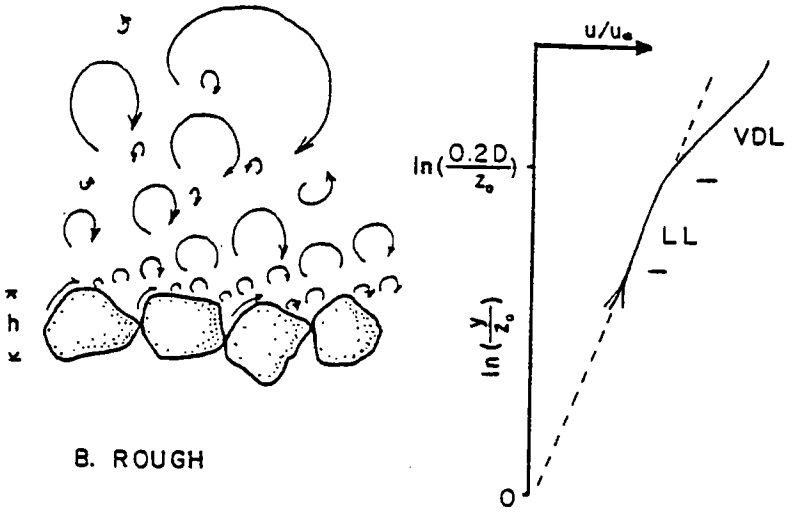
VSL - Viscous sublayer

ILL - Integrated logarithmic layer

SL - Surface layer



A. SMOOTH



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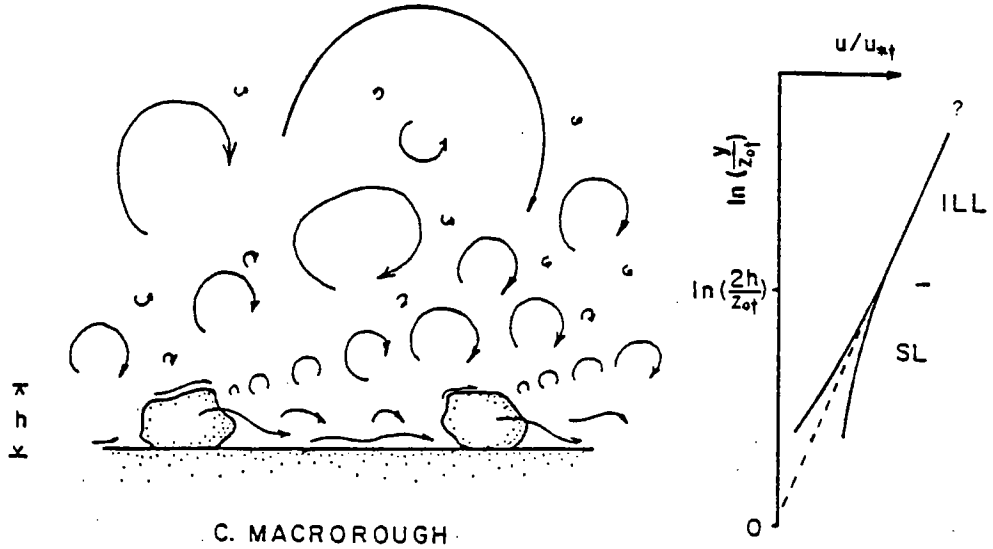

C. MACROROUGH

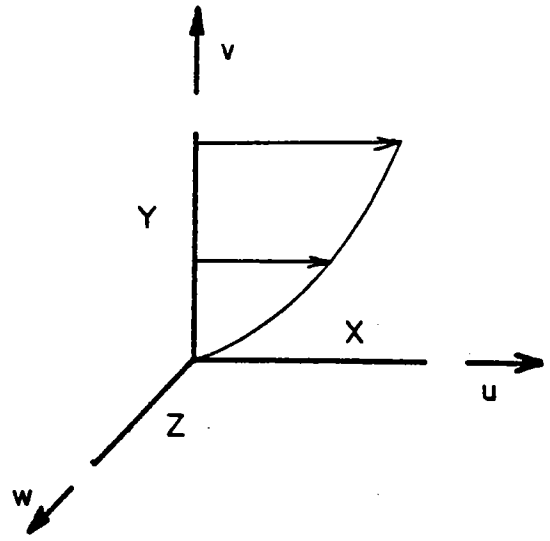

Figure 1.2. General Cartesian coordinate system (x,y,z) and corresponding velocity components (u,v,w).

(usually gravity or pressure), it must be balanced by the boundary shear stress. Hence for the outer layer

$$\frac{u_s - u}{u*} = g(\frac{y}{D}) \qquad 1.2$$

Millikan (1939) showed that by requiring that both (1.1) and (1.2) hold in some "overlap region" the velocity profile there is specified to within two empirical constants:

$$\frac{u}{u*} = a_1 \ln(\frac{u*y}{\nu}) + a_2 \qquad 1.3$$

This relation was originally derived in a different way, so to maintain consistency a_1 is usually written as $1/\kappa$ where κ is called the von Karman constant. Much has been written about the constancy of κ; it certainly has the value 0.4 in smooth flow for the three general types of boundary layer named above over the range of Reynolds numbers attainable in the laboratory. This makes (1.3) a very powerful tool: it is relatively difficult to measure u* directly, particularly in field experiments, but it can be determined from (1.3) as $u* = \kappa du/d(\ln(y))$. Making the required velocity-profile measurements is usually relatively straightforward.

The numerical value of a_2 is of less interest for our purposes. It is similarly constant at about 5.0.

So for a smooth turbulent boundary layer the velocity profile looks as shown in figure 1.1a. Immediately above the bed there is a viscous-dominated region called the viscous sublayer in which the profile is linear: $(u/u*) = (u*y/\nu)$. This is overlain by a buffer or transitional region and then

the logarithmic layer; the velocity profile there is often loosely referred to as the "law of the wall" although strictly speaking this governs the two layers below as well. Beyond this is the outer or "velocity-defect" region, governed by a form of (1.2) that varies according to whether the boundary layer is fully developed in a pipe or in an open channel, or is not fully developed.

Now consider a turbulent boundary layer over a uniform bed of coarse sand of diameter h (figure 1.1b). By analogy with flow about an isolated sphere one would expect that, if velocities near the bed are large enough, flow separation should occur around the grains and the resulting wakes should disrupt the velocity field. We define a roughness Reynolds number $R* \equiv u*h/\nu$; empirically, if R* is greater than about 5, the viscous sublayer is affected by the presence of the grains, and if R* is greater than about 70 (fully rough flow) it is completely destroyed in a formal sense. Of course, there must still be a viscous-dominated region on the surface of each grain to satisfy the no-slip condition, and with it a viscous shear stress. Continuing our analogy with the behavior of an isolated sphere, however, we expect the dominant part of the drag on the grains to be pressure or form drag induced by the separated flow about the grains. The boundary shear stress, in which this drag is averaged over the bed, is thus independent of viscosity and so of R*.

The scaling arguments outlined for smooth flow can be extended to the fully rough case, provided we exclude from consideration the region very near the bed where the grains affect the flow field individually (figure 1.1b). The length scale in the law of the wall includes the viscosity, which is clearly inappropriate here; it is natural instead to choose h, the grain height, as the inner length scale. We retain u_* as the scaling velocity and write the law of the wall as

$$\frac{u}{u_*} = f_r\left(\frac{y}{h}\right) \qquad 1.4$$

The outer flow, on the other hand, includes no explicit dependence on ν so there is no reason to modify (1.2) for rough flows. In an overlap region where (1.4) and (1.2) are both valid, we obtain the rough-bed equivalent to (1.3):

$$\frac{u}{u_*} = a_3 \ln\left(\frac{y}{h}\right) + a_4 \qquad 1.5$$

It is one of the most fundamental and remarkable results of the study of turbulent boundary layers that $a_3 = a_1$; that is, that the relation between u_* and $du/d(\ln(y))$ in the logarithmic region is the same in both smooth and rough flows. On the other hand, a_4 is different from the smooth case and depends on the geometry of the roughness. It is usual to rewrite (1.5) as

$$\frac{u}{u_*} = \frac{1}{\kappa}\ln\left(\frac{y}{z_o}\right) \qquad 1.6$$

where z_o is called the roughness length; it is a length scale proportional to the scale of the roughness. The constant of proportionality depends on the roughness geometry. (Here I

am generalizing slightly from the example of closely packed sand under consideration to similar kinds of small-scale, uniformly distributed roughness.) For closely packed sand-grain roughness, $z_o=h/30$.

It is possible to imagine sediment transport under either smooth or rough conditions as described above, although for the smooth case the viscous sublayer is disrupted by the moving grains (Gust and Southard, in press) and the length scale in R* may need to be redefined. In any case the force propelling the sediment is the boundary shear stress $\rho u*^2$: for both kinds of boundary layer considered so far, this is the average tangential force per unit area on the grains.

Now suppose we have a similar situation -- a bed roughened with large (say a few centimeters), closely packed spherical grains -- but imagine them to be laid in a single layer on a flat, smooth surface. Water flows over the bed so that u* is about 1 cm/s and R* is a few hundred, comfortably above the limit for fully rough flow. What happens if we remove grains one at a time, leaving individual grains surrounded by smooth surface (figure 1.1c)? (It would be necessary to adjust the mean velocity continuously to keep the boundary shear stress constant.) The large grains still separate the flow, exert form resistance, and shed wakes that locally disrupt the viscous sublayer. But the no-slip condition implies the existence of a viscous-dominated region near the wall: at the bed surface the boundary shear stress has the same

viscous form as it does on a smooth wall: $\tau_o = \rho\nu(\partial u_p/\partial y_n)_o$, where y_n is locally normal to the bed and has its origin at the bed surface, and u_p is the velocity locally parallel to the surface. Such small-scale, local shear stress is known as skin friction. Due to gradients in pressure and turbulence intensity induced by the presence of the large grains, the skin friction varies with position over the bed.

We may define the total bottom stress τ_{ot} as

$$\tau_{ot} = \frac{1}{A}\left(\int_{A'} p_o n_y dA' + \int_{A'} \rho\nu\left(\frac{\partial u_p}{\partial y_n}\right)_o n_y dA'\right) \qquad 1.7$$

where A is an averaging area that includes a number of roughness elements, A' is a true wetted surface area, p_o is the local pressure on the boundary, n_y is the vertical component of a unit vector normal to the surface, and u_p is the velocity parallel to the surface. This is the total boundary shear stress acting against the flow, so that in a wide, uniform open channel of depth D and slope S,

$$\tau_{ot} = \rho g D S \qquad 1.8$$

The first term in (1.7) is called the <u>form drag</u> (τ_{of}); it is meaningful only in a spatially averaged sense. On the other hand, the <u>skin friction</u> (τ_{os}) is the integrand in the second term; it is a well-defined local quantity although it enters the total bottom stress in spatially integrated form.

The boundary layer over a bed of widely and perhaps irregularly spaced roughness elements is thus considerably more complicated than either a rough or a smooth boundary layer; it really combines elements of both. The work to be

described in this thesis is aimed at clarifying aspects of this type of flow, which will be referred to as macrorough. Most of the further discussion of macrorough flow will be left to later, more detailed sections, but a few additional general comments should be made here. First, each of the three boundary shear stresses in (1.7) -- τ_{ot}, τ_{of}, and τ_{os} -- can be converted to a friction velocity $u* = (\tau_o)^{1/2}$. In light of the role of $u*$ in scaling velocity profiles, embodied in (1.2), (1.3) and (1.6), one would expect the composite nature of the boundary shear stress to be reflected in the velocity field. We can get a rough idea of how this works by considering the consequences of a general increase in the length scale of the turbulence with height, which in turn is a simple kinematic result of the increasing distance of the eddy centers from the wall (Townsend, 1976, p. 156 ff.). As eddies become larger, they respond more slowly to changes in strain rate, and as their distances from the wall increase, they are advected more quickly by the mean velocity; the net result of both effects is that the area of the bed to which the turbulence responds increases with height in the flow (Townsend, 1965a). (These ideas will be developed in more detail in section 3.53.) Since the total bottom stress τ_{ot} is spatially averaged by definition (1.7), it can be a valid scaling parameter only above some height at which the turbulence is large enough to average the variable bottom stress. This height cannot be less than some small

distance above the tops of the roughness elements (empirically, it will be seen in section 2.32 to be about one roughness height), because the form drag, included in (1.7) as an integral of pressure on the boundary, appears in the flow as excess Reynolds stress generated in the wakes of the roughness elements. One must be somewhere above the tops of the roughness elements before the wakes will have merged to produce a Reynolds-stress field that is uniform in the streamwise direction. Below this, the velocity and turbulence fields vary spatially in response to the rough topography.

Combining all of the above, the following picture emerges. In macrorough flow, the spatially integrated region well above the tops of the roughness elements corresponds to the logarithmic layer given by (1.6); it will be referred to here as the integrated logarithmic layer (ILL). In deriving the profile law (1.6) for fully rough flow, a single friction velocity equivalent to $u*_t$ emerged naturally as the velocity scale because there was no need to consider the details of the viscous shear stress on the grain surfaces. But in macrorough flow there are areas between roughness elements where the boundary shear stress is purely viscous; furthermore the disposition of this viscous stress and its contribution to the total boundary shear stress both depend on the geometry and arrangement of the roughness elements.