Loose-boundary hydraulics and fluid mechanics: selected advances since 1961

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SUMMARY: A major and acknowledged role in the advance of sedimentological knowledge is played by the methods of physics as expressed through fluid mechanics and loose-boundary hydraulics. Turbulence is increasingly being seen as involving orderly flow structures, and these are significant for the origin of several sedimentary structures and for suspension transport. Recent advances in sediment-transport theory are rooted in the concept of the sediment load as a downward-acting force and in the notion of the flow as a transporting machine. Laboratory experiments are increasing our understanding of bedforms due to rivers and waves, but there is a need for well-framed field studies to clarify the influence of change in natural environments. Physical explanations are being proposed for tidal bedforms. Field studies in both modern environments and the stratigraphic record suggest that tidal patterns and strengths are recognizable from the internal features of cross-bedded units formed by sand waves. Tidal patterns are complex, however, and a wider range of situations needs to be explored than has hitherto been the case. Mathematical modelling could help to define the expectable bedding patterns. Although successful mathematical models have been constructed of flow and sedimentation in channel bends, secondary flow in real bends and the evolution of those bends are not yet well understood. Turbidity currents on a geologically significant scale have not been observed directly, but valuable insights into their character and processes have come from laboratory experiments. Recent laboratory work points to the limitations of lock-exchange experiments and emphasizes the various ways in which turbidity currents mix with the ambient medium. These studies seem to have implications, yet to be worked out, for the internal features and some sole marks of turbidites.

In 1961 the British Sedimentology Research Group stood on the threshold of a period of explosive growth in sedimentological activity in the British Isles, during which the number of those claiming to be sedimentologists has grown by an order of magnitude. From a minor topic of doubtful import on the geological horizon, sedimentology has matured into a discipline with a central place in the geological curriculum and a powerful and acknowledged role in the advance of geology's principles and applications. Moreover, sedimentology has come to be recognized as multidisciplinary in the best sense, for in it we find strands from many traditional geological fields interwoven with threads from several fields of non-geological scholarship. The cloth would have appealed to the Renaissance mind.

Some of us who stood at that threshold many years ago saw that a valid contribution to sedimentology was to be made from the fields of loose-boundary hydraulics and fluid mechanics, where the experimental and theoretical methods of physics are applied to the problems of the motion of pure fluids and of fluid–solid mixtures. We hoped thereby to achieve a better understanding of sediment erosion, transport and deposition, in all their varied natural modes, and a fuller appreciation of sedimentary structures and their physical and environmental meanings. Our vision was not new, for Sorby (1859, 1908) a century previously had demonstrated the power of a quantitative and experimental approach to sediments, but we did feel that now its time had come.

The extent to which the methods of physics have been applied to an understanding of sedimentary processes and products has increased at much the same pace as interest in sedimentology as a whole. There is now hardly a physical process of sedimentation which has not been illuminated by laboratory experiments, quantitative studies in modern environments, and mathematical explorations starting from general principles. Indeed, the wealth of data available today is so great that the reviewer with limited space faces insuperable problems of choice. The following personal selection of topics—turbulent flow, sediment transport, bedforms, flow in channel bends, and turbidity currents—is certainly not comprehensive, but it perhaps serves to illustrate the advances that have been made in several important areas. In most it is possible to see the interaction of well-framed field studies, experiment and theory.

Turbulent boundary layers

Turbulence—the apparently random motion of a fluid which is commonly also in translation—is a phenomenon of global significance. Most natural flows in the atmosphere or hydrosphere are turbulent, and turbulence is critically important to a wide range of industrial and engineering processes. Natural turbulence arises chiefly where a fluid flows and therefore shears past a stationary boundary—e.g. a river in its bed—or where the flow separates at a bluff body—e.g. a boulder projecting up into the current. However, turbulence also arises, especially in the ocean and the atmosphere, both of which are stratified, at the interface between two extensive layers of fluid in relative motion.

Early students of turbulence were impressed by its randomness, and therefore adopted an essentially statistical-dynamical approach to turbulent flows. To the considerable understanding of turbulence and its effects achieved by these means (see Townsend 1976 for review) has been added in the last 20 years a growing appreciation of the fact that turbulence is also to a considerable degree orderly (see Cantwell 1981 for review). By using various techniques, particularly flow visualization, it has been found that turbulent currents are largely formed of flow configurations of a definite and reproducible size, shape and persistence in time, i.e. structural features that are coherent. Moreover, these configurations seem to be hierarchical and their temporal and spatial scales can be related to flow properties. Turbulence is therefore not just a form of chaos, but a strongly ordered phenomenon in which stochastic and deterministic processes and effects are married together. The coherent structures of turbulence are best understood from boundary-layer flows.

The turbulent boundary layer is divisible into at least two regions: a wall region where the flow in the innermost part (viscous sublayer) is essentially laminar and where elsewhere vortical structures and turbulence begin to develop, and an outer region where the turbulence completes its development and predominates. Kline *et al.* (1967) used visualization and hot-wire techniques to show that the fluid in the wall region, and especially in the viscous sublayer, was organized into an array of streamwise high- and low-speed streaks whose transverse spacing depended only on the boundary shear stress and fluid properties (see also Smith & Metzler, 1983). Periodically at a fixed point, the fluid in part of a high-speed streak became lifted up and swiftly ejected on a steeply inclined path into the outer flow, a process called streak bursting. Later work (see particularly Grass 1971), stimulated by this important finding, has modified and refined the concept of streak-bursting and our notion of the coherent structure represented by a bursting streak (see reviews by Bridge 1978a; Cantwell 1981; Allen 1982a). The work particularly of Grass (1971), Utami & Ueno (1977) and of Head & Bandyopadhyay (1981) strongly suggests that the bursting streak is a jet of low-momentum fluid ejected from near the wall into the outer part of the flow in the form of a horseshoe vortex (modification of axisymmetric jet by mean shear flow). At low Reynolds numbers these obliquely inclined vortices are loop-shaped but at large numbers are greatly stretched axially into a hairpin-like form. Even far out into the flow these vortices retain the transverse scale of the low-speed streaks from which they appear to be derived (Head & Bandyopadhyay 1981). Significantly, the frequency of streak-bursting is determined not by wall parameters like the boundary shear stress but by the overall flow properties, notably thickness and mean velocity (Rao *et al.* 1971). However, there is probably no universally constant non-dimensional burst frequency (Bandyopadhyay 1982).

The measurement of the motion of hydrogen bubbles generated in one flow-visualization technique (Grass 1971), and the clever analysis of hot-wire signals (Brodkey *et al.* 1974), have uncovered some important dynamical properties of turbulent wall regions filled with bursting streaks. Contrary to widespread belief, a marked anisotropy, particularly normal to the wall, characterizes the turbulence. The fluctuating velocity components directed perpendicularly away from the wall prove to be significantly stronger than those directed inward toward the boundary. Therefore there is a substantial net outward flow of momentum from the wall into the outer region of the turbulent boundary layer.

Early work on turbulent boundary layers and other shear flows established that their outermost edges are only intermittently turbulent and suggested that such flows comprise an array of large eddies or vortices on the scale of the boundary layer itself (Townsend 1956). More recent visualization and hot-wire studies, notably those of Kovasznay *et al.* (1970), Blackwelder & Kovasznay (1972), Brown & Thomas (1977), Falco (1977), Head & Bandyopadhyay (1981), and Thomas & Bull (1983) confirmed this result and showed how the horseshoe vortices representing bursting streaks could be related to these larger vortical structures, which may also be horseshoe-shaped. The large eddies, also inclined obliquely to the flow boundary but at a shallower angle, are two to three boundary-layer thicknesses long and about one boundary-layer thickness wide. They travel with the flow at approximately the mean

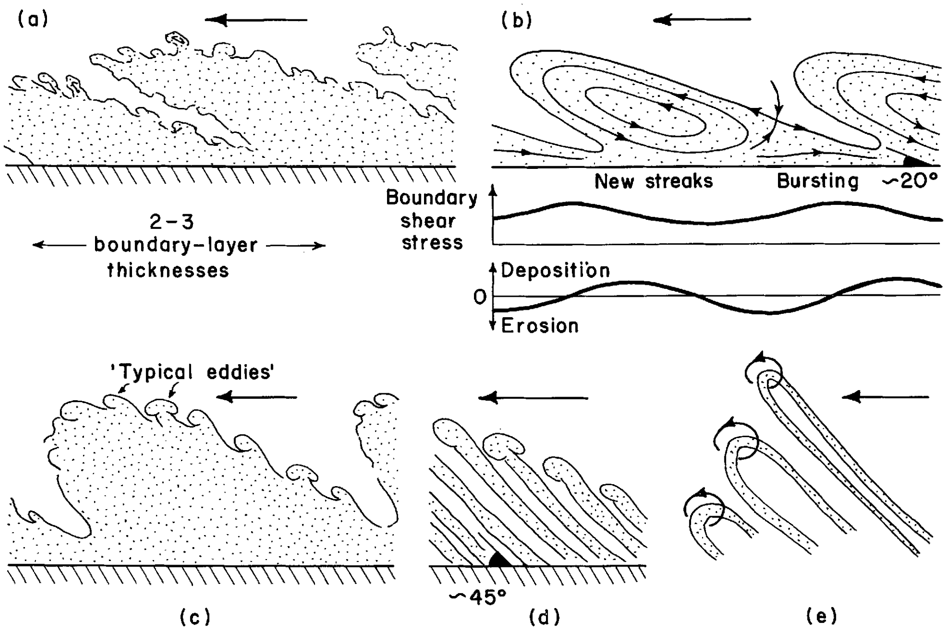

FIG. 1. Summary of experimental evidence for coherent structures in two-dimensional turbulent boundary layers on plane walls. (a) Position of the interface between marked (stippled) and unmarked fluid showing the occurrence of large-scale structures (possibly horseshoe-shaped) on the scale of the boundary layer itself (adapted from Falco 1977). (b) Flow pattern within the large structures as seen by an observer travelling approximately with the structures (adapted from Brown & Thomas 1977 and Falco 1977). The streamwise distribution of boundary shear stress is also shown, together with its implications for sediment bedload transport erosion and deposition. (c) The tips of vortices due to bursting streaks ('typical eddies' of Falco 1977) expressed on the upper surface of a large-scale structure. (d) Stack of vortices due to bursting streaks within the proximal part of a large-scale structure (adapted from Head & Bandyopadhyay 1981). (e) Form of vortices due to bursting streaks at (from left to right) small, intermediate and large Reynolds numbers (adapted from Head & Bandyopadhyay 1981).

flow speed. In a wall-related turbulent flow (Brown & Thomas 1977; Thomas & Bull 1983), the boundary shear stress is a maximum (and also strongly fluctuating) at or just upstream of the gaps between consecutive large eddies. Each large eddy appears to consist of numerous horsehoe vortices representing burst streaks, the tops of which would seem to be Falco's (1977) 'typical eddies'.

Figure 1 is an attempt using this work to portray a model of the coherent structures and related properties of turbulent boundary layers, with each part of the diagram emphasizing a particular aspect. It seems possible that streak-bursting is controlled by the movement past the bed of the large horseshoe vortices of the outer flow (Brown & Thomas 1977). The streamlines in these vortices are concave-outward near the wall, a feature suggesting that the wall streaks represent a Görtler-type instability maintained by the passage of one large eddy after another. The bursting of the streaks will clearly be favoured by the rise in boundary shear stress observed toward the rear of each large horseshoe vortex. Streak-bursting at a fixed point will occur at about the observed frequency, if the vortices are considered to be arranged *en echelon* in rows across the flow. Nakagawa & Nezu (1981) found that the streamwise spatial separation of bursts forming at the same instant was two to three flow thicknesses.

Sedimentologists concerned with bedforms quickly seized on these important advances in fluid mechanics. Parting lineations are widely familiar structures from parallel-laminated sands and sandstones, and it is now clear that these streamwise ridges and furrows reflect boundary-layer streaks and streak-bursting (Allen 1970a; Mantz 1978) (see also reviews by Bridge 1978a; Allen 1982a). Bridge (1978a) suggests that the laminations themselves record streak-bursting. However, individual laminae, carrying multitudinous lineations, are far too extensive across the flow to be due to the bursting of single streaks (see Head & Bandyopadhyay, 1981), but their scale and other characters are consistent with the downstream convection of the large horseshoe vortices, beneath which there is a streamwise variation in boundary shear stress (Brown & Thomas 1977) appropriate to a cyclical pattern of short-term erosion and deposition (Fig. 1). Jackson (1976a) attributed dunes in sand-bed rivers to streak-bursting, on the basis of measurements of the frequency with which large eddies appeared at fixed points on the surface of the Wabash River. These eddies are far too big to be individual burst streaks (see Head & Bandyopadhyay, 1981) and are probably to be identified either with the large horseshoe vortices of laboratory-scale turbulent boundary layers, or as vortices generated in the free-shear zones of separated flows coupled to the dunes. Insofar as streaks and their bursting appear to depend on the properties and convection of the large vortical structures (Brown & Thomas 1977), it seems unlikely that streak-bursting of itself can be the cause of dunes.

Sediment transport

General

There can be no detrital sedimentary record without sediment particle erosion, transport and deposition. Specifically, there can be no record unless the sediment transport rate varies in space and/or time, and in such a manner that, under tectonic, eustatic or topographic influences, deposition in the long term prevails over erosion. An understanding of sediment transport, and particularly its space–time variability on comparatively small scales, is therefore crucial to an understanding of sedimentary sequences.

Bagnold's physically based theory of sediment transport is unquestionably the most significant single contribution of recent decades to this area of sedimentology. The theory was advanced in a long and difficult paper (Bagnold 1956) that failed to catch widespread attention. However, the later more accessible and somewhat widened version (Bagnold 1966) has profoundly influenced many areas of sedimentological thought. Refinement of the theory by its author continues to the present day (Bagnold 1973, 1977, 1979, 1980).

The essence of Bagnold's theory is two-fold. Because the common rock-forming minerals are more dense than either water or air, detritus maintained against gravity in a state of transport above the lower boundary of a current is a load pressing down on the bed, no matter what the mode of transport. This load is equal to the immersed weight of the particles per unit bed area or, alternatively, to the quotient of the sediment transport rate (immersed weight basis) and the mean grain transport velocity. But a load pressing down on the bed must under conditions of uniform steady transport be supported by an equal but upward-acting force. The ultimate source of this force can only be the transporting agent itself. The second essence lies in the fact that the rate of sediment transport has the quality and dimensions of a rate of doing work. Therefore a sediment-transporting flow can be regarded as a machine, the transport rate being linked through an efficiency factor to the power of the flow, i.e. its potential ability to do work. Hence for such as rivers, the wind, or turbidity currents flowing over a static bed, the transport rate must be proportional through a non-dimensional coefficient to the product of a characteristic boundary shear stress and a characteristic flow velocity. Alternatively, the transport rate may be written as proportional through dimensional coefficients, either to the cube of a characteristic flow velocity, or to the 3/2nd power of a characteristic boundary shear-stress term.

The flow-derived force supporting the load is divisible into two distinct components, each relating to a specific dynamically defined mode of sediment transport (Bagnold 1966) (see also Leeder 1979a, b). The bedload under practical conditions comprises relatively large particles in dense array confined to near the bed, making frequent contacts with the bed and with each other, and which are supported above the bed by particle impact forces. On the other hand, the suspended load under practical conditions comprises relatively small particles at a low concentration distributed throughout the whole flow and which thus seldom, if ever, touch. These particles would seem to be buoyed up by turbulence. Bagnold postulated that wall-related turbulence is anistropic in a direction normal to the boundary, so that the upward fluctuations in a stream exceed the downward ones and thus yield the net upward momentum flux necessary to support the suspended grains against gravity.

Bedload

Bagnold's (1956, 1966) bedload-transport theory rests on his study of the impact forces arising during the bulk shearing of neutrally buoyant uniform waxy spheres dispersed in a fluid confined in the annular space between two coaxial rotating drums. Curiously, few later attempts have been made either to reproduce or extend these rather idealized experiments. Savage (1978, 1979) and Savage & Jeffrey (1981) took some steps in this direction, partly confirming Bagnold's results, and there have been encouraging parallel developments on the theoretical side (Shen & Ackermann 1982). But there is a pressing need for further work, particularly that aimed at clarifying the role of particle size, shape and density as affecting the bedload transport of real grains.

Much more attention has been given both experimentally and theoretically to individual bedload particles travelling over a granular bed, especially when the particles are transported in the absence of moving neighbours. Early Japanese work in this area is an important source of data that has long been overlooked (Tsuchiya 1969, 1971; Tsuchiya *et al.* 1969; Tsuchiya & Aoyama 1970; Tsuchiya & Kawata 1971, 1973). Later contributions were made by Gordon *et al.* (1972), Ellwood *et al.* (1975), Luque & Van Beek (1976), White & Schulz (1977), Reizes (1978), Murphy & Hooshiari (1982) and, particularly, Abbott & Francis (1977). These studies show that a bedload grain may either slide, roll or saltate over the bed, the incidence of saltation relative to the other two modes increasing as the driving force exceeds the threshold for grain entrainment. Bedload grains in water take leaps between successive bed contacts that are of the general order of one to three grain diameters high and ten grain diameters long. The trajectories depend strongly on spin and added mass, as well as on the impact and driving forces. Grains saltating beneath the wind take trajectories generally many orders of magnitude larger than the same particles advancing in water. Forces due to spin continue to be important.

The bedload transport rate depends critically on the conditions for grain entrainment as well as on the applied force. Whereas threshold conditions are well established for practically horizontal beds under a wide range of conditions (Komar & Miller 1973, 1975a; Miller & Komar 1977; Miller *et al.* 1977; Dingler 1979), the influence of appreciable bed slope on the transport rate is virtually unknown quantitatively. Only Lysne (1969) explored slope effects experimentally, yet they appear to be crucial to the maintenance and limiting shape of many bedforms (Bagnold 1956; Fredsøe 1974; Richards 1980; Allen 1982b, c, d). A smaller entrainment force is needed for downslope than upslope transport, whence a bedform tends to decay unless a counterbalancing mechanism also operates. Bed slope has an additional effect, as yet unexplored except theoretically, namely, on

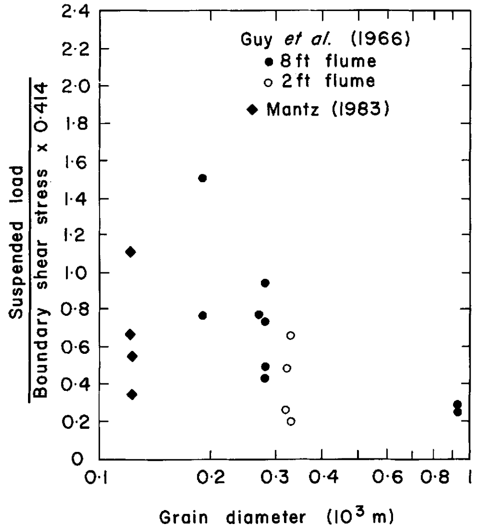

FIG. 2. Relative suspended load measured in laboratory experiments above upper-stage plane sand beds as a function of grain size (data of Guy *et al.* 1966, Tables 2–7; Mantz 1983, Tables 3, 6 and 9).

the transport rate itself (Bagnold 1956). More work is necessary for upslope than downslope transport, because in transport up a slope the load must be lifted against gravity as well as maintained above the static bed. There is an urgent need to explore these slope effects further.

Suspension

Suspension transport has traditionally been associated with fluid turbulence but treated as a diffusion phenomenon. Classically, the downward flux of suspended grains due to gravity is balanced against an opposite flux along an upward-declining concentration gradient (Rouse 1937). This analysis suffers from the weakness of being purely kinematic; it is impossible to deduce either the magnitude of the suspended load or of the suspension transport rate. Furthermore, the single function traditionally used to describe the vertical concentration of suspended sediment serves to conceal important differences in variation in different regions of the flow. In an inner region, the concentration declines according to an inverse linear relationship, but in the outer region decays logarithmically (Coleman 1969, 1970).

An interesting recent development is McTigue's (1981) demonstration that statements derived from the diffusion model can be regained using the *dynamical* theory of mixtures, in which the solid phase is allowed volume and weight. Essentially, an equilibrium can be struck between the immersed weight of the suspended particles and the fluid drag on them due to turbulence. The dynamical implications of mixture theory for suspension transport should now be explored.

Bagnold (1966) also accepted a connection between suspension transport and turbulence, but pursued an apparently very different dynamical analysis. He pointed out that turbulence could only support the suspended load if it was anisotropic, in the manner already discussed, and argued from a knowledge of shear turbulence that the load (immersed weight) should equal approximately 41% of the mean boundary shear stress. These proposals were widely discounted at the time, despite Irmay's (1960) independent evidence adduced in their favour. A major change of view has since occurred. The experiments of Grass (1971) and Brodkey *et al.* (1974) show that wall-related turbulence is anisotropic precisely as required. Moreover, as Leeder (1983) pointed out, experimentally determined suspended sediment loads above plane sand beds (Guy *et al.* 1966; Mantz 1983) are of the same order as the fraction of the mean boundary shear stress calculated by Bagnold (Fig. 2). The grain-size effect suggested by the data may either depend on the apparatus used or reflect the way the dispersed grains modify the turbulence through their drag and added mass.

We seem therefore to be approaching a period of renewed integration in sediment transport research. Suspension transport seems to depend on the anisotropy of shear turbulence, which in turn is one expression of the ordered character of a turbulent boundary layer. The experimental studies of Sumer & Ogoz (1978) and of Sumer & Deigaard (1981) draw attention to the fact that the coherent structures now recognized as occurring in all turbulent flows must take an increasingly important place in future work on the mechanisms of suspension transport. Coleman's (1969, 1970) inner and outer sediment concentration regions should have an explanation in these coherent structures and the turbulence associated with them.

Bedforms and their internal structures

Unidirectional aqueous currents

The sandy bedforms generated by one-way water streams furnish through their external shapes and internal geometry major clues to the meaning of the sedimentary record. From them can be inferred the direction and intensity of sediment transport and, when allied with other evidence, a semi-quantitative indication of flow conditions. Although Owens (1908) and Gilbert (1914) had long ago glimpsed the existence of a succession of bedforms with increasing flow strength, a reasonably detailed and comprehensive hydraulic description of this sequence has only recently arisen, largely through engineering laboratory experiments made at the Colorado State University (Fort Collins) (Guy *et al.* 1966). Simons *et al.* (1965) first drew wide attention to this work amongst geologists.

The relationship of bedforms to flow under steady uniform conditions in straight channels of laboratory scale can be expressed from the very considerable experimental data now available in stability or existence diagrams of several types. Figure 3 shows bedforms in the unit stream power–grain-size plane (Simons *et al.* 1965; Allen 1968, 1970, 1982a). The velocity–depth–grain-size scheme of Southard (1971) is also popular (Southard & Boguchwal 1973; Costello & Southard 1981). Bedforms found in natural environments, which are non-uniform and unsteady, and generally much deeper than laboratory flumes, are fairly satisfactorily described using the stream power–grain-size scheme (Smith 1971; Bridge & Jarvis 1976, 1982) and the velocity–depth–grain-size plot (Boothroyd & Hubbard 1974; Jackson 1976b; Dalrymple *et al.* 1978; Cant 1978; Rubin & McCulloch 1980). However, the comparisons are so far limited to comparatively small-scale systems. In a third and less-often-used scheme, bedforms appear in a graph of boundary shear stress (either dimensional or non-dimensional) against grain size (e.g. Leeder 1980; Allen 1982a). A fourth scheme of limited use shows bedforms in the shear-stress–grain-Reynolds-number plane (e.g. Allen & Leeder 1980).

Each scheme is restricted in applicability by the limitations of the data base (Fig. 3). Mantz (1978, 1980) has gone some way toward improving knowledge of bedforms in silts, but much is unknown about sediments of this important size class. Bedforms in gravels are also poorly understood and some workers appear to think that the larger sandy bedforms have no coarser-grade counterparts. The field occurrence of breakout-flood gravel dunes (e.g. Baker 1973) and of gravel antidunes (Shaw & Kellerhals 1977) warns against this prejudice. For all size classes, however, there is a dearth of observations at large stream powers. Do upper-stage plane beds persist to an indefinitely large stream power (assuming no free-surface effects), or can further instabilities develop? Finally, the schemes apply only to near-spherical quartz

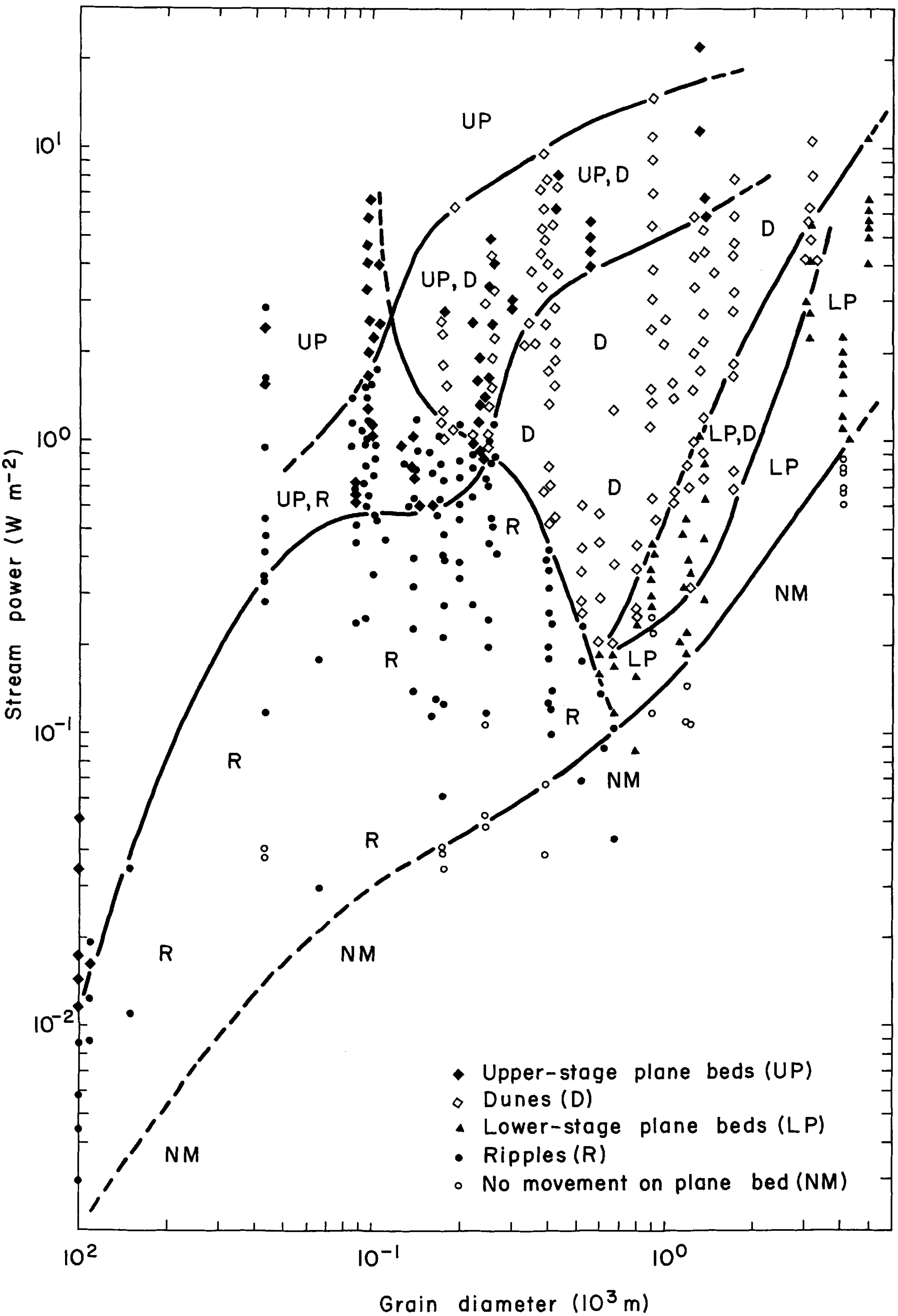

FIG. 3. Existence of bedforms in uniform-steady flow in straight laboratory channels as depicted in the stream power–grain-size plane, based on 566 observations from nineteen investigations (for data sources see Allen 1982a, p. 339). To avoid overcrowding only representative and critical data points are shown. The stream power is calculated using wall-corrected values for the boundary shear stress, and the grain diameter is adjusted to plain water at 25°C.

sands; virtually nothing is known of the flow relationships of bedforms composed of strongly non-spherical particles—e.g. shell hash.

Velocity–depth–grain-size schemes have the particular disadvantage of becoming increasingly unreliable as the natural systems to which they are applied grow larger relative to the laboratory scale. The schemes involving stream power and boundary shear stress have the limitation of showing the fields for plane beds significantly overlapped by those for ripples and dunes (Fig. 3), for which only part of the total flow force is directly involved in the sediment transport. Bridge (1981) explained the overlap between the dune and upper-stage plane bed fields, but that with lower-stage plane beds remains mysterious.

Insofar as they depend on steady-state experiments, none of the schemes is applicable without considerable caution to the sedimentary record, shaped by unsteady and non-uniform flows. Allen (1973, 1974) and Allen & Collinson (1974) pointed out that bedforms, especially large ones, should lag in size and shape behind changes of flow, because sediment transport can occur only at a finite, flow-determined rate. There is now abundant evidence for lag, both from experiments (Simons & Richardson 1962; Gee 1975; Wijbenga & Klaasen 1983) and the field (Pretious & Blench 1951; Allen *et al.* 1969; Stückrath 1969; Peters 1971; Nasner 1974). Two theoretically possible mechanisms of bedform response to a changing flow (Allen 1976a) are (i) alteration of the composition of the bedform population, as new individuals better adjusted to the changed conditions replace existing forms, and (ii) change on the part of individual forms during their lifespans. The operation of these or other as yet unidentified mechanisms leads to complex and non-unique relationships between bedform and flow properties (e.g. Allen 1976b, c, 1978a), which somewhat restricts our

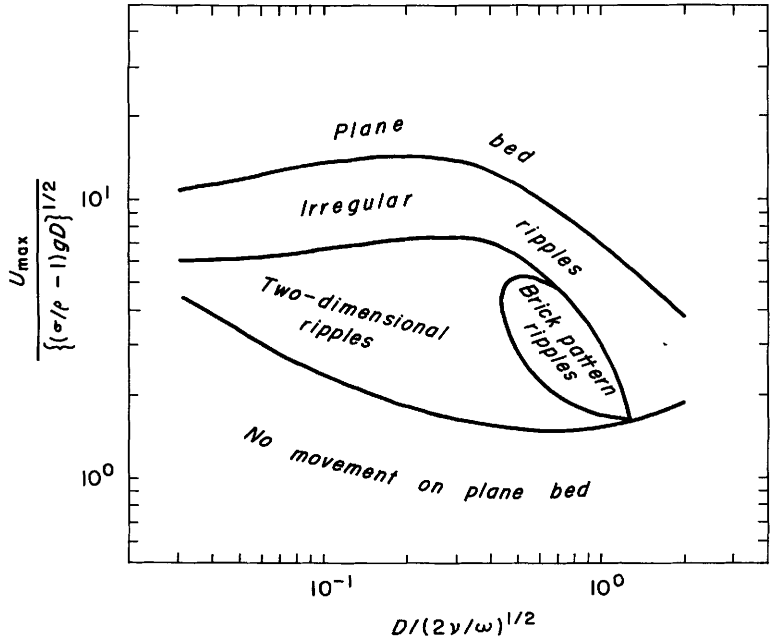

FIG. 4. Existence of bedforms related to water waves, based on experiments using monochromatic waves in a straight wave channel (after Kaneko 1980). The ordinate is a non-dimensional boundary shear stress, where U_{max} is the maximum orbital velocity of a near-bed water particle, σ and ρ respectively the solids and fluid densities, g the acceleration due to gravity, and D the sediment particle diameter. The abscissa shows a non-dimensional grain size (denominator is the thickness of the wave boundary layer), where ν is the fluid kinematic viscosity and ω the angular frequency of the waves.

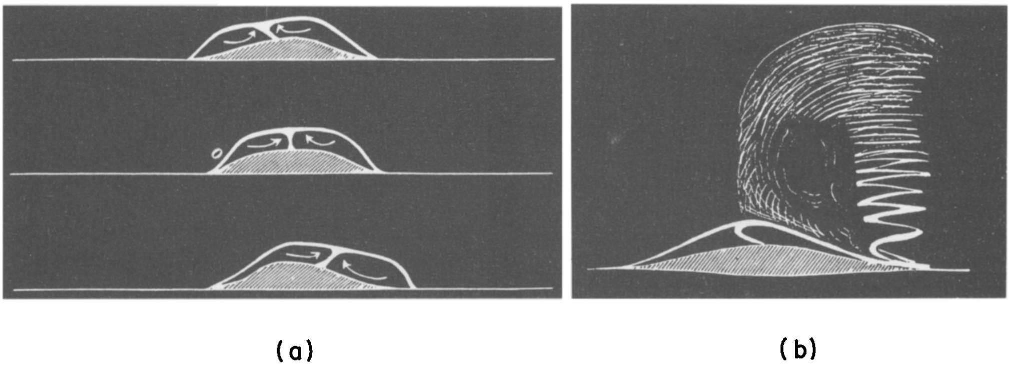

FIG. 5. Reproductions of two of Darwin's (1884) original figures to illustrate the two-layer drift currents generated by waves acting on deformable granular beds. (a) The 'ink mushroom' at different stages in the wave cycle, illustrating the lower layer of vortices, the drift just above the bed being from trough to crest. (b) The 'ink tree' illustrating the upper layer of vortices, in which the drift is upward in the trough and downward above the crest (an ink mushroom appears below).

ability to interpret bedforms and their internal structures. However, lag does permit transient bedforms to be preserved—e.g. dune forms in flash-flood deposits.

Attempts by mathematical modelling to understand lag mechanisms and their effects were made by Allen (1976b, d, e, 1978a, b) and Fredsøe (1979, 1981). Allen's model relates to strongly unsteady flows, whereas Fredsøe's is more restricted in terms of the mechanisms incorporated, and is at present suitable only for gradually changing flows. Both models give results in qualitative agreement with field data. Neither model incorporates the effects of flow non-uniformity, an important limitation in comparison with natural environments, and each could be improved if more were known experimentally about lag. What exactly are the mechanisms of lag and what is their quantitative significance?

Existence diagrams such as Fig. 3 describe but do not explain sandy bedforms in one-way currents. The physically most appealing explanation of the bedforms is that they record either the stability (plane bed) or instability (wavy bed) of a deformable granular boundary in the presence of a grain-transporting flow. Kennedy (1963, 1964, 1969) made the first comprehensive stability analysis of bedforms; the approach is now a powerful and sophisticated tool, although still with limitations. Kennedy showed mathematically that bed waves (ripples, dunes, antidunes) could grow only where there was an appropriate spatial lag between the streamwise changes of bed elevation and the streamwise changes in the rate of sediment transport over the bed. Thus a bed wave can grow in amplitude only when the transport-rate maximum lies upstream of the wave crest. Damping occurs when the crest lies upstream. However, Kennedy's models for the flow and transport were very simple, and his lag distance was introduced as an arbitrary parameter. An intrinsic spatial lag emerges from later analyses using more realistic flow and transport models (Engelund 1970; Fredsøe 1974; Parker 1975; Nakagawa & Tsujimoto 1980; Richards 1980). The lag distance may depend on as many as six distinct effects, which in each flow combine algebraically to determine its magnitude (Allen 1983). Some reflect the way the fluid alone responds to the wavy bed, and others the response of the bedload and suspended load to the spatially changing current. These later analyses are quite good at predicting bedform existence fields and draw attention to the importance of bed slope and sediment transport mode as controls on bed stability. Because non-linear effects are ignored, they predict the quantitative aspects of bedforms less well. Puls (1981) and Fredsøe (1982) have recently and with some success calculated dune size and shape.

Currents generated by wind waves

After long neglect, attention has again been turned to the bedforms generated by wave-related currents, particularly the ripple marks. Figure 4 shows the existence field for these bedforms suggested experimentally by Kaneko (1980), others having been proposed by Allen (1967, 1970a) and Komar & Miller (1975b). Based on laboratory and field data, several models for the interpretation of ancient wave ripple marks have now been advanced (Komar 1974; Clifton 1976; Allen 1979a; Allen 1981).

Perhaps the most important development has been the rediscovery and substantial vindication through extensive theoretical and experimental studies of Darwin's (1884) original explanation for wave ripple marks. Using ink to mark the fluid, Darwin found that the oscillatory current present above a bed of the ripples (not too steep) was accompanied by slow drifts of fluid which defined a vertically stacked stationary double circulation. The drift created what he called his ink tree and ink mushroom, as reproduced in Fig. 5 from his original paper. The lower row of stationary recirculating vortices, denoted by the ink mushrooms, is very flat and the near-bed flow is from ripple troughs to crests. The opposite circulation, giving the ink trees, is observed in the upper row of vortices, which are distant from the bed and much larger. The response of the boundary to the near-bed drift was a movement of grains from ripple trough to crest and a consequent increase in the amplitude of the bed waves and further accentuation of the drift. The bed would appear to have been intrinsically unstable, which is not surprising since no natural granular boundary can be without drift-inducing waviness.

Darwin's (1884) flow pattern was not again seen experimentally until Uda & Hino (1975). Physically, it depends on force gradients that arise in an oscillatory boundary-layer flow that thickens and thins over bed waviness. Several theoretical predictions of these drift currents exist (Lyne 1971; Hall 1974; Sleath 1974, 1976; Uda & Hino 1975; Kaneko & Honji 1979; Matsunaga *et al.* 1981), supported by abundant experimental proof (Uda & Hino 1975; Hino & Fujisaki 1977; Kaneko & Honji 1979; Honji *et al.* 1980; Matsunaga & Honji 1980; Du Toit & Sleath 1981).

When wave ripple marks become sufficiently steep, effects due to flow separation seem to dominate and the forms become attributable to Bagnold's (1946) vortex type. Tunstall & Inman (1975) and Longuet-Higgins (1981) recently explored the flow over vortex ripples but confined themselves to the separation-related vortices. Darwin's drifts are present (Du Toit & Sleath 1981) but their influence in maintaining the bed features is as yet uncertain.

Tidal currents

Oscillatory tidal currents generate a variety of bedforms, the general morphology and distribution of which are now well known (Stride 1982). The mechanics of the larger of these—sand ribbons, sand waves (excluding dunes as understood from one-way flows) and tidal current ridges—have attracted some recent attention, although there is so far little agreement as to interpretation.

Sand ribbons are longitudinal bedforms that are most commonly seen as streamwise bands of sand on a surface of contrasted texture. It has long been held that these structures reflect the action of a pattern of secondary flow in the form of paired streamwise corkscrew vortices. McLean (1981) emphasized the role of transverse roughness variations in the inducement and maintenance of such vortices, and the dependence of their scale on flow thickness. However, there are other possible causes—e.g. turbulence anisotropy and wave–wind drift interactions (Langmuir vortices). What is lacking from the field is a study designed to confirm the presence of a secondary flow in association with sand ribbons and to establish its most probable cause. There is also considerable uncertainty about the scale of sand ribbons. Their spacing as summarized by Allen (1982e) using published field data increases on the average with flow thickness, and in a manner consistent with what is theoretically expected of secondary flows composed of paired streamwise vortices. Kenyon (1970) and Belderson *et al.* (1982) believe that ribbon spacing is unrelated to flow depth.

Tidal current ridges (Off 1963) are very large linear sand banks roughly aligned with tidal currents that occur about equally spaced in groups. Numerous sand waves lie superimposed on them. Unlike sand ribbons, with a spacing typically a few times the water depth, tidal current ridges lie a distance apart approximately two orders of magnitude greater than the depth (Off 1963; Allen 1968). The ridge crests are offset by about 10° from the direction of the regional peak tidal streams (Smith 1969; McCave 1979; Kenyon *et al.* 1981), and most of the banks have markedly asymmetrical cross-sectional profiles.

An origin by the action of secondary flows similar to those responsible for sand ribbons (Off 1963; Houbolt 1968) is supported neither by the relative spacing of the ridges nor perhaps by McCave's (1979) field observations of the local currents. The large relative spacing probably also discounts their dependence implied by Kenyon *et al.* (1981) on the centrifugal instability of the tidal boundary layer. Caston (1972) concluded from field observations that the strongest tidal currents and sand transport tended on the shallower parts of a bank to turn toward the crest, maintaining the form of the bank and giving a net sand transport in the direction of the steeper face (see also

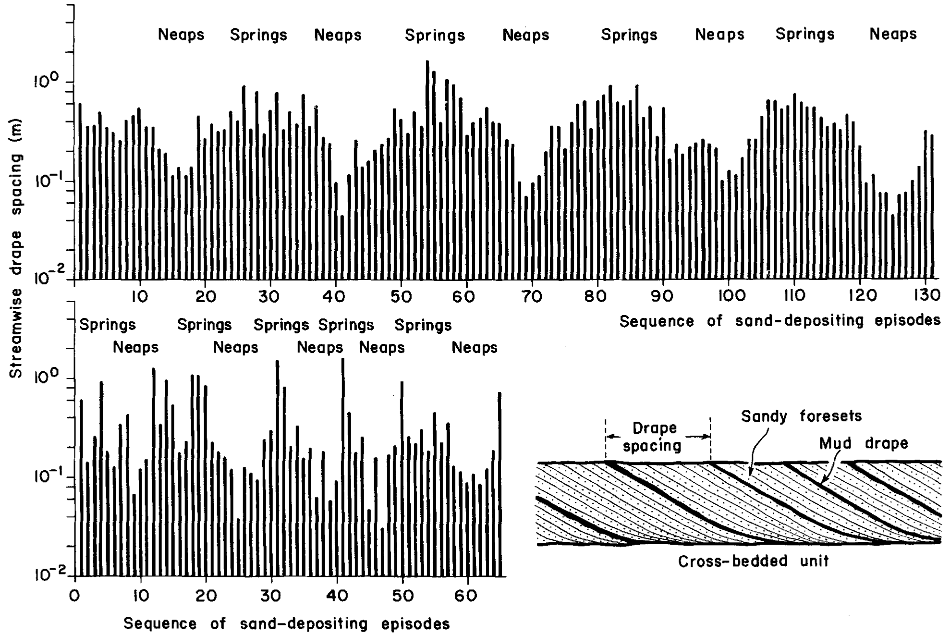

FIG. 6. Graphs illustrating spatial patterns in the arrangement of mud drapes in tidally formed cross-bedded units. The upper graph represents a late Holocene unit from the Oosterschelde (data of Visser 1980), and the lower a unit from the Folkestone Beds (Cretaceous) in the English Weald (data of Allen 1981, 1982f).

Caston & Stride 1970). In addition, there appears to be a circulation of sand around a bank (Houbolt 1968; McCave & Langhorne 1982), best revealed by the orientations of sand waves superimposed on the sand ridges. Huthnance (1973, 1982a, b) has given an important and quantitatively correct analytical justification of Caston's (1972) model, the banks being interpreted as due to an instability involving the influence of bed slope on sediment-transport rate.

Tidal current ridges are expected to comprise cross-bedded sands organized within a complex hierarchical framework of erosional bedding contacts (McCave & Langhorne 1982; Stride *et al.* 1982). Because the bank itself is apparently migrating in one direction, but sand is circulating around it, only the internal master accretion bedding would appear to denote the direction of regional net sand transport. The internally preserved cross-bedding related to dunes superimposed on sand waves, and any intermediate-order structures attributable to the sand waves themselves (Allen 1980), are those due to the oppositely acting subordinate tidal stream. Care should therefore be taken when interpreting palaeocurrent patterns from shallow-marine sandstones.

Sand waves ranging from symmetrical to asymmetrical in cross-sectional profile, and typically with dunes superimposed on them, are the largest transverse bedforms shaped by tidal currents. Their cause is as obscure now as at the time of their discovery. Some workers (Cartwright 1959; Furnes 1974) have invoked internal waves to explain them, but there are good reasons why this explanation cannot be universal. Allen (1979b) advanced the possibility that sand waves are analagous in the oscillatory tidal flows to the drift-related ripples generated by surface waves. Belderson *et al.* (1982) saw sand waves merely as modified dunes but did not satisfactorily explain what seem to be the dunes superimposed on them. Hammond & Heathershaw (1981) advanced a general wave theory which, when a beat effect is included, predicts the scale of sand waves well. The work of Davies (1982a, b) and Heathershaw (1982) on the reflection of wave energy by sea-bed topography may help to explain some sand waves. Theoretically, sand waves grow in asymmetry with increasing net sediment transport (Allen 1982c, d), and critical field studies are now needed to amplify and refine the few existing observations (Allen 1980) linking sand-wave shape with the flow and sediment-transport regimes.

Finally, we urgently need to establish by direct observation the internal structure of sand waves, and how in detail that structure is related to bedform shape, size and flow-transport regime. Models based on shallow sampling (e.g. Reineck 1963), and on considerations of form, movement and general regime (McCave 1971; Allen 1980; Langhorne 1982), will be helpful for a time, but cannot be regarded as ultimately satisfactory. Recent studies (e.g. Visser 1980) have stressed the importance of slack-water mud drapes and reactivation structures preserved within cross-bedded units as a means of unravelling tidal flow patterns, particularly the spring–neap cycle. As examples, Fig. 6 shows the downcurrent spacing of mud drapes in a subfossil cross-bedded deposit from a Dutch tidal channel (Visser 1980) and in a cross-bedded unit from the Cretaceous Folkestone Beds of the Weald (Allen 1981, 1982f). The closely spaced drapes suggest neap tides too weak to transport much sand, and those spaced further apart the more vigorous springs. The pattern of spacings in each case is roughly periodic, the length of the cycles suggesting in the one a diurnal and in the other a semidiurnal tidal regime. There are suggestions in Visser's (1980) case of the diurnal inequality widely known from European tides. Valuable as are the recent field investigations of Boersma (1969), Terwindt (1971, 1975, 1981), Visser (1980), Boersma & Terwindt (1981), Kohsiek & Terwindt (1981) and Van den Berg (1982), it should be remembered that they relate to a mesotidal to macrotidal and strongly semidiurnal regime. Tidal bedding patterns recorded from the stratigraphic record (Levell 1980; Allen 1981, 1982f; Homewood 1981; Homewood & Allen 1981) suggest that other tidal regimes should be studied before general conclusions are drawn regarding structural expressions of tidal cyclicity. Tidal regimes are complex and varied. Mathematical modelling could help to define the kinds of bedding patterns to be expected under different regimes.

Channel bends

Water flowing in an open, curved, rigid laboratory channel of moderate width/depth ratio follows a simple corkscrew path, drifting toward the outside of the bend just beneath the free surface, and inward near the bed. This type of secondary motion is due to a viscosity-related imbalance between the outward-directed centrifugal and inward-acting pressure forces on each fluid element. Sand introduced into the flow accumulates on the insides of the bends, in response to the inward-directed component of the bed shear stress (Yen 1970; Martvall & Nilsson 1972; Onishi *et al.* 1972; Hooke 1975; Kikkawa *et al.* 1976; Zimmerman & Kennedy 1978). When the channel boundaries are altered from rigid to mobile, as in Ackers & Charlton's (1970) experiments, erosion on the outside of each bend proceeds in harmony with accretion on the inner bank, and the bend migrates sideways and downstream. Under some circumstances, a limiting bend amplitude is reached, after which the migration is downstream only.

It is not surprising that the theory and actuality of secondary flow in channel bends, and its implications for sediment transport, bend behaviour, and fluvial landscapes, should have long attracted the attention of hydraulic engineers, geomorphologists and sedimentologists. The history of research in this field is cautionary, however, for it offers a particularly clear instance of how advances can be delayed when workers operating within different traditions proceed in ignorance of each other's activities (Allen 1978c). Van Bendegom (1947), an engineer working in the Netherlands, gave the first substantial flow-based analysis of bend sedimentation, but because he published in Dutch his advance was widely overlooked. Much later, and unknown to each other, at least two groups of hydraulic engineers and a sedimentologist arrived at much the same quantitative model for deposition in channel bends.

Only the invincibly ignorant can any longer deny the relevance of secondary flow to the sedimentary processes and behaviour of curved river and tidal channels. Nonetheless, although field data overwhelmingly prove the occurrence of secondary flow (Hey & Thorne 1975; Jackson 1975; Bridge & Jarvis 1976, 1977, 1982; Bathurst *et al.* 1977, 1979; Hickin 1978; Dietrich *et al.* 1979; Thorne & Hey 1979), the flow patterns observed seldom resemble the single corkscrews generated in the laboratory. Commonly, several vortices are present, forming a sheaf whose strands loosely twist and untwist, variously tilt, and flatten and narrow as they are traced through a bend or series of bends. A good example is afforded by the River South Esk, the data of Bridge & Jarvis (1982) being re-analysed in Fig. 7. Similar vertically stacked vortices are occasionally reported from the laboratory (Prus-Chacinski 1954; Mosonyi *et al.* 1975). Moreover, as Bathurst *et al.* (1979) also observe, the number and pattern of the vortices can change substantially as the plan and cross-sectional shape of the bend vary in response to discharge fluctuations (Fig. 7). Indeed, at some river stages, the handedness of the secondary flow dominating a bend may be the opposite of that implied by the bend curvature (e.g. Jackson

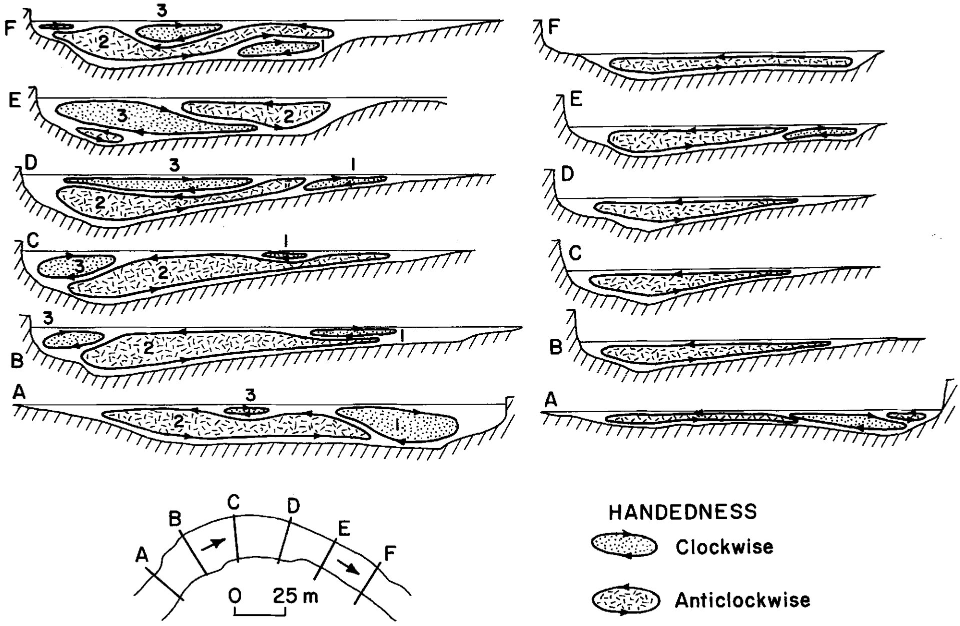

FIG. 7. Direction of rotation (observer looks downstream) of secondary flows in a single meander bend of the River South Esk, Scotland. Left-hand set of profiles for discharge of 19.6 $m^3 s^{-1}$, and right-hand for discharge of 4.3 $m^3 s^{-1}$. A tentative correlation of the vortices from section to section is suggested by the numbering applied at the higher discharge value. Based on a re-analysis of data originally published by Bridge & Jarvis (1982).

1975 as analysed by Allen 1982e; Hickin 1978). At which point value of the discharge is a given bend best adjusted with reference to the secondary flow developed? Are all the bends in a given stretch of a river best adjusted to the same discharge? Clearly, much more remains to be learned from the field about secondary flows and their influence. The problem in the tidal case has scarcely been touched.

In view of the complexity of secondary flow in real channels, it is perhaps surprising that attempts to model sedimentation in channel bends should have been so successful. Two kinds of mathematical model exist. The less useful (e.g. Suga 1967; Odgaard 1982) seeks to establish from dynamical considerations the channel shape and distribution of sediment, given the static stability of the bed material. The other, outlined by Van Bendegom (1947), and subsequently elaborated in two dimensions (Allen 1970a, b; Yen 1970; Ikeda 1974, 1975; Bridge 1976; Kikkawa *et al.* 1976) and three dimensions (Engelund 1974; Gottlieb 1976; Bridge 1977 1978b; Zimmermann & Kennedy 1978; Bridge & Jarvis 1982), seeks the condition for the dynamical stability of sediment in transport as bedload. In essence, dynamical stability exists when, at every point on the bed, the upslope component of the fluid force acting on the total bedload or a single bedload particle is exactly balanced by the downslope-acting sediment weight. Combined with the use of bedform existence diagrams (e.g. Fig. 3), this type of analysis yields a remarkably accurate picture of the horizontal and vertical variation of grain size and sedimentary structures in river point bar deposits (Bridge 1978; Bridge & Jarvis 1982). But there remain some outstanding questions. Is Van Bendegom's (1947) type of force-balance equation preferable to Engelund's (1974)? Are we justified with Rozovskii (1961) in regarding the strength of secondary flow in channel bends as effectively independent of bed roughness (see Zimmermann 1977)?

Morphological and historical evidence widely suggests that river channel bends typically migrate both downstream and laterally. We are far from understanding how channel bends develop in this way, but one possible mechanism involves a spatial lag between the channel shape in plan and the spatially changing erosive potential of the secondary flow determined by that shape (Allen 1974, 1982e). On a more local scale, Begin (1981) argues that the same interaction (a lag is not however involved) explains a supposed limiting shape for channel

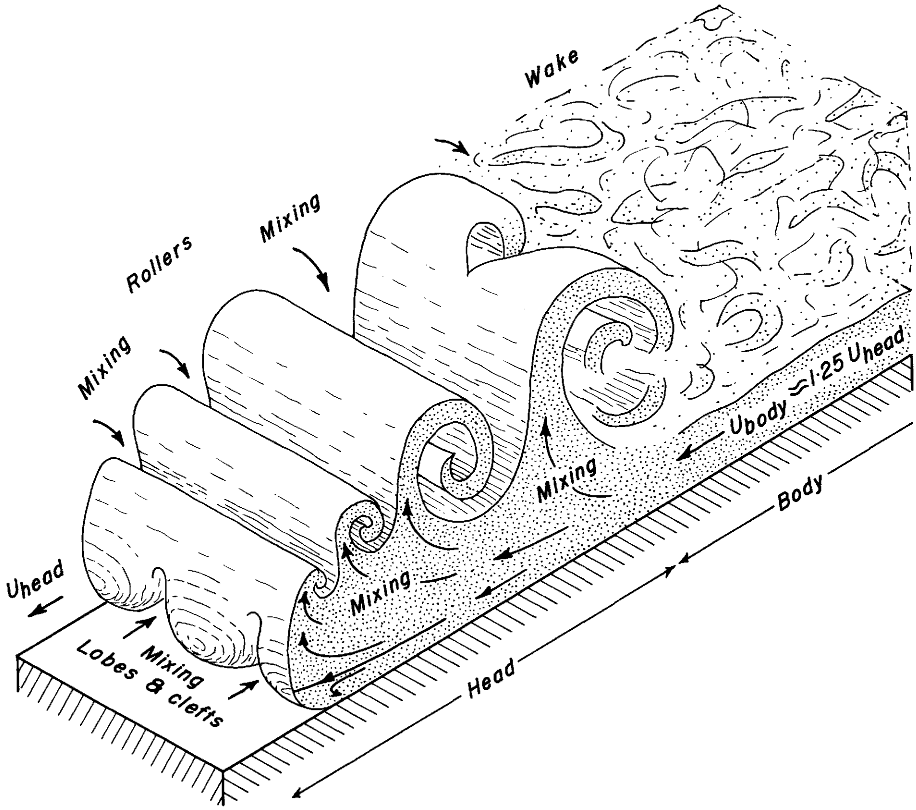

FIG. 8. Suggested structure of the leading portion of a turbidity current advancing in deep water over a nearly horizontal bottom. Based chiefly on Simpson (1969, 1972), Allen (1971), and Simpson & Britter (1979).

bends. The theory of Ikeda *et al.* (1981) and Parker *et al.* (1982) treats meandering as an instability phenomenon and attributes no effective role to either secondary flow or lag. Two instability mechanisms operating on similar characteristic wavelengths are identified; one is related to bar deposition and the other to outer-bank erosion. The theory gives good agreement with observation and is particularly appealing in that some non-linear effects observed from natural channel patterns are reproduced. The next step with all these theories must be to make numerical models capable of predicting the initiation of meanders in straight channels and their growth to realistically large amplitudes. What is the factor limiting amplitude growth in the absence of constraining bedrock walls?

Turbidity currents

Turbidity currents belong to the important class of flows called gravity currents (see Chen 1980 and Simpson 1982 for reviews). These are flows driven by a *small* difference of density relative to the neighbouring fluid or fluids. Depending on relative density, a gravity current may flow either beneath or over the surface of the ambient medium, or at some level within it when stratified. Turbidity currents are driven by an excess of density due to the presence of dispersed sediment particles.

Kuenen & Migliorini's (1950) revolutionary paper persuaded geologists that turbidity currents must be treated seriously as agents for the transport of large quantities of coarse sediment from shallow to deep water. The acceptance of the hypothesis, however, did not spark off amongst geologists quite the experimental and theoretical activity that might have been expected, although there are mitigating circumstances. Turbidity currents belong to a class of time-dependent, non-linear, non-uniform and unsteady free-boundary flows (Chen 1980), and are further complicated by containing as an essential element particles capable of sedimenting under gravity. The experiments of Middleton (1966a, b, 1967), Lovell (1971), Kersey & Hsu (1976) and Lüthi (1981a, b) provided valuable insights into the flow and transport of sediment by turbidity currents, but have the now-evident serious limitation of heavy reliance on lock-exchange techniques with few if any natural parallels. On the theoretical front, Bagnold (1962) emphasized the idea that turbidity currents might derive their energy directly from a part or the whole of the sediment dispersed in them, namely, that they are autosuspended. Middleton (1966c) and Southard & Mackintosh (1981) disputed this concept, but others have given it their support (Pantin 1979; Parker 1982). These largely untested studies highlight the fact that the key to modelling turbidity currents is how to marry the behaviour of sediment-free gravity currents with a proper treatment of the sediment; a part of this may contribute to autosuspension, another portion may be fully supported in transport by the available fluid forces, while the third and probably substantially greater part exceeds the supportable load and so is available for progressive deposition.

Probably the most important and immediately applicable advances in this difficult area have come from carefully married theoretical and experimental attempts to understand atmospheric gravity currents (Simpson 1969, 1972; Britter & Simpson 1978, 1981; Simpson & Britter 1979, 1980; Britter & Linden 1980). These suggest that turbidity currents in deep water have the general form and motion summarized in Fig. 8. The head of the current is overhanging, as has long been known, and is divided transversely into roughly periodic buttock-shaped lobes and clefts (Simpson 1969, 1972). These structures record the viscous nature of the ambient medium into which the current is advancing and, in particular, the gravitational (Rayleigh-Taylor) instability that arises where the less dense medium is being over-ridden by the heavier current (Allen 1971). The clefts carry narrow streams of ambient medium back into the head and so permit a small amount of mixing into the current (Allen 1971; Simpson 1972; Simpson & Britter 1979). Associated with the lobe and cleft structure should be a transverse variation in boundary shear stress which could explain some of the longitudinal sole markings of turbidites (Allen 1971, 1982e). Simpson (1969, 1972) further showed that spatially growing transverse billows were generated on the top of the head. These billows, representing Kelvin-Helmholtz instability, record a high rate of mixing into the medium from the current (Britter & Simpson 1978; Simpson & Britter 1979). On account of the mixing, the body of a current in deep water flows about 25% faster than the head, so that the current consumes itself in the process of mixing into the ambient medium. Indeed, the mixing and instability of the head is the crucial factor in determining the overall motion, whether on a horizontal or sloping bed (Simpson & Britter 1979; Britter & Linden 1980).

The self-consuming nature of gravity currents emphasized in Simpson & Britter's (1979) work raises the interesting possibility that turbidity currents in a large-scale oceanic setting may develop episodically. As has often been suggested, one may envisage a typical turbidity current as generated high up on a continental margin by the mixing and dilution of slumps and slides with sea water. The current will be of finite volume and therefore length in the canyons and valleys through which at first it flows. The consumption of the last of the body by the head as the current flows along signals the completion of the creation as a kind of wake of a much larger volume of fluid which, because it contains some dispersed sediment, will be capable of further downslope flow. This process of consumption and transformation through mixing into a larger and more dilute but still finite mass might occur more than once during the flow of a turbidity current over an extensive ocean floor. Could the grain-size break at the 'bedding joints' of turbidites (e.g. Parkash 1970) express an episode in such a cascading development?

Conclusion

The advances made since 1961 in our understanding of turbulence, sediment transport, bedforms, flow in channel bends, and turbidity currents convincingly demonstrate the general value of applying the methods of physics in its broad sense to sedimentological problems. Well-designed quantitative field observations allow us some opportunity to link process and product under natural conditions, but these conditions are generally so complicated that few of the linkages can be established fully, even when many examples have been made available. Laboratory experiments allow us to test hypotheses and to explore processes under controlled and reproducible circumstances. But they are good only to the extent that we know which factors are critical under natural conditions, and suffer particularly from limitations of scale. Mathematical modelling is a powerful means for gaining insight into the controls on natural processes and especially into the relative importance of the factors involved; the value of quantitative statements about the natural process cannot be overestimated. At a more developed level, mathematical models can reproduce with considerable accuracy the behaviour and outcomes of particular natural sedimentary systems, but at present only when these are comparatively simple. There have therefore been solid gains which future generations of sedimentologists will consolidate and build upon. More power (so to speak) to them!

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