3

Fluvial processes

The morphology of natural river channels is determined by the interaction of the fluid flow with the erodible materials in the channel boundary. The underlying problem is to understand that interaction given that it involves the distinct processes of entrainment, transport and deposition of sediment. The basic mechanical principles are well established but a complete analytical solution is still a long way off, largely because natural streams represent the movement of a fluid–solid mixture in boundaries that are themselves deformable. Even the motion of a single particle cannot be described adequately and the problem becomes more complex when the boundary material is cohesive. To the extent that a larger flow implies a greater force, a discussion of streamflow fluctuation provides a background to the mechanical work performed by streams.

Streamflow fluctuation

Water reaching a stream channel may have followed any one of the routes identified previously (Figure 2.6A). Each route gives a different response to rainfall or snowmelt in terms of the volume of flow produced and the timing of contributions to the channel. Consequently the character of discharge variation is determined by the relative contribution from each source, which is in turn influenced by climatic factors and the physical characteristics of a basin.

For convenience total flow is traditionally subdivided into two parts: direct or storm runoff, and baseflow (Figure 3.1). The distinction is based on the time of arrival in the stream rather than the route followed, although the two are necessarily related. Direct runoff consists of surface flow (Horton or saturation overland flow) and a substantial part of the subsurface flow, while baseflow is mainly supplied from groundwater sources. Since water flowing at depth beneath the surface moves relatively slowly, its outflow into the stream lags behind the occurrence of rainfall and tends to be very regular. Groundwater flow thus maintains a steady basal flow throughout the year in more humid environments.

Most measurements of stream discharge are made at gauging stations, where discharge (Q) is defined by the continuity equation as the product of cross-sectional area (A = w.d) and mean velocity (v)

$$Q = A.v = w.d.v \tag{3.1}$$

Provided a long enough record exists, measurements can be manipulated to yield a flow-duration curve which shows the frequency with which discharges of different magnitude are equalled or exceeded. Thus a discharge of 7.6 $m^3 s^{-1}$ is equalled or exceeded about 10 per cent of the time at the Bollin gauging station (Figure 3.2A). Mean annual discharge generally has a frequency of about 25 per cent and occupies approximately 40 per cent of the total capacity (bankfull cross-sectional area) of

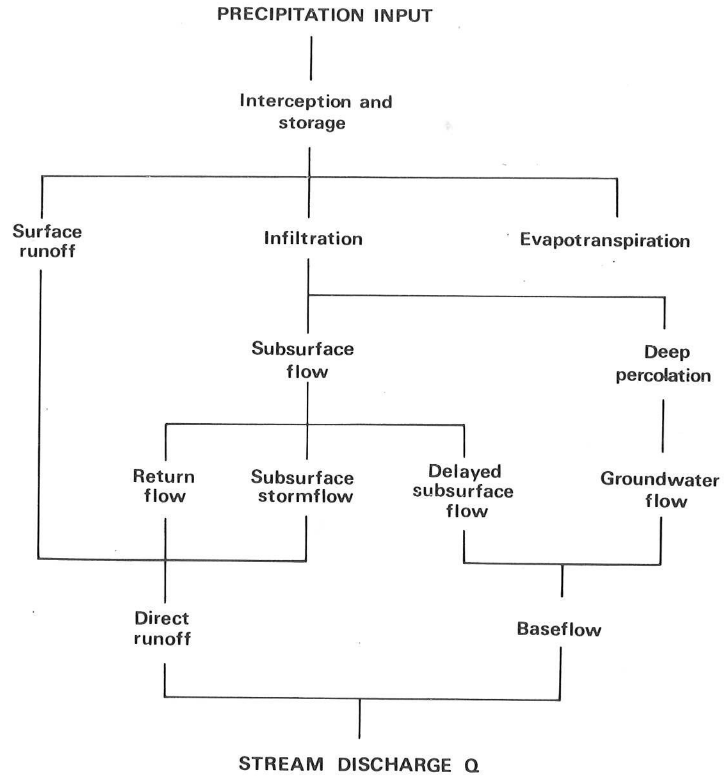


Fig 3.1 Schematic diagram of the runoff process.

the channel (Dunne and Leopold, 1978). If a sediment rating curve relating transport rate to discharge is also available, the flow-duration curve can be converted into a cumulative sediment transport curve to show the contributions made by various discharges to the total load transported.

Various procedures exist for computing flood frequencies. One of the simplest is based on discharge maxima in a series of years, from which the recurrence interval or return period (T in years) of flood events can be calculated:

$$T = (n + 1)/N \tag{3.2}$$

where n is the number of years of record and N is the rank of a particular event. The resultant flood-frequency curve shows the average time interval within which a flood of given size will occur as an annual maximum (Figure 3.2B). Thus in any one year the probability of the annual maximum exceeding a flood with a recurrence interval of 10 years is $\frac{1}{10}$. The mean annual flood with a return period of about 2.33 years ($Q_{2.33}$) has occasionally been used in relating channel form variables to discharge (e.g. Ferguson, 1973). Another flow with assumed morphologic significance is bankfull discharge (Q_b), defined as that discharge at which the channel is completely full, with a reported recurrence interval of 1.5 years in the United States (Leopold *et al.*, 1964). These kinds of data are useful for regional analyses of flow characteristics and for classifying the river regimes of moderately sized basins. The

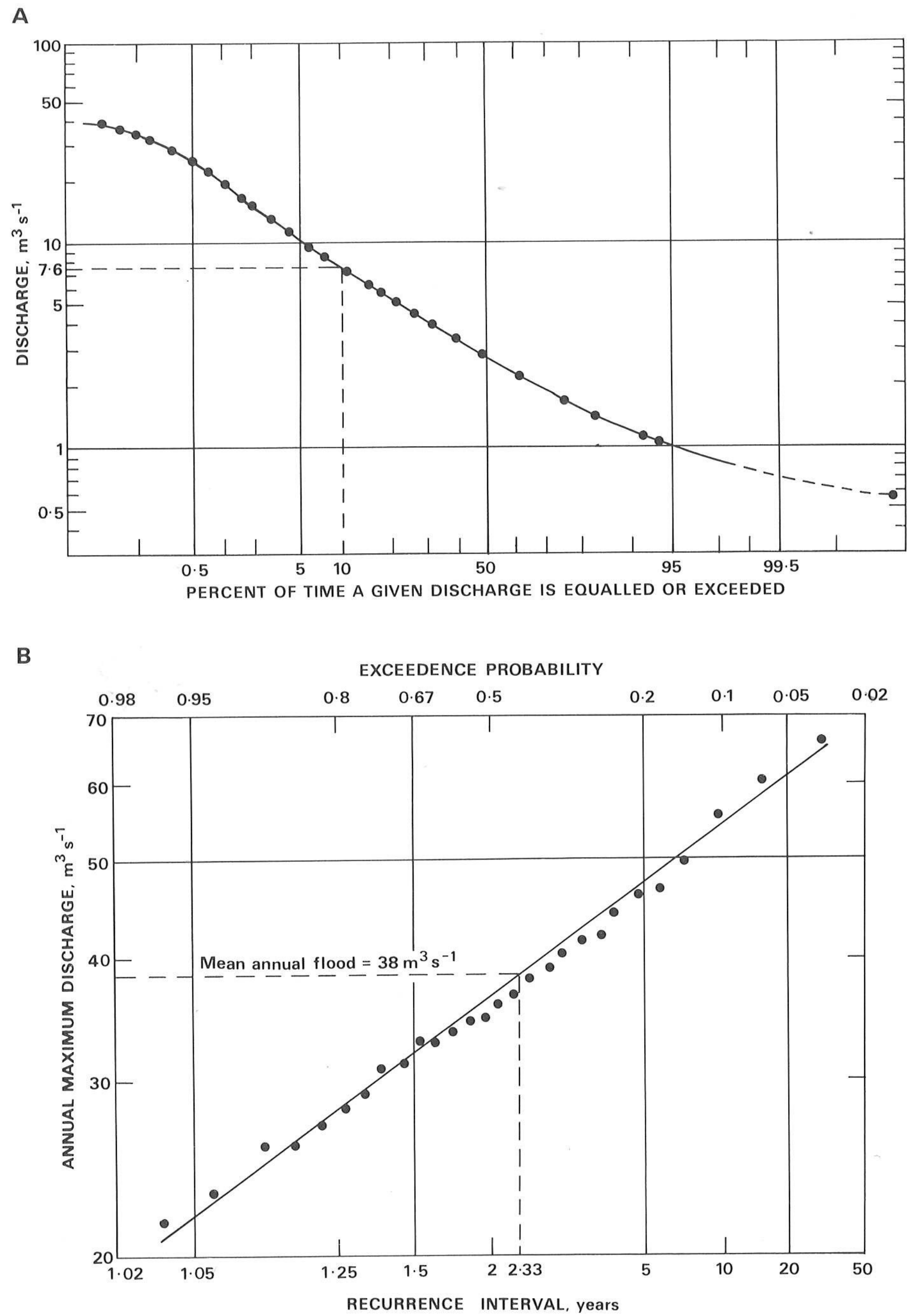


Fig 3.2 Flow-duration (A) and flood-frequency (B) curves for the River Bollin, Cheshire. The top scale of the flood-frequency curve gives the probability that the corresponding discharge is equalled or exceeded in any one year, while the bottom scale gives the average number of years in which the annual peak equals or exceeds the corresponding discharge.

regime of a river may have an important influence on channel form adjustment (Harvey, 1969; Stevens *et al.*, 1975).

Not surprisingly discharge is highly correlated with drainage area (Figure 1.2A). In many basins discharge of a given frequency (f) increases less rapidly than drainage area (A_d) to give an exponent n in

$$Q_f = cA_d^{\,n} \tag{3.3}$$

which is less than 1. The value of n is not independent of the frequency of flow considered or drainage area, tending to increase with more frequent flows and decrease with increasing basin size because of storage effects. Since drainage area is easier to measure, it has often been used as a surrogate for discharge in empirical studies of channel morphology.

The shape of the discharge hydrograph at any point depends on the way in which water is added to the stream and on the storage characteristics of the basin. Runoff supplied to a channel moves downstream as a wave of increasing and then decreasing discharge. The transmission of a flood wave is subject to two main effects which alter its character:

(i) translation, in which the wave moves downstream without any significant change in shape; and

(ii) reservoir action, whereby the time base of the wave is lengthened through temporary storage in the channel and valley bottom.

In upstream parts where drainage area is small and slopes are steep, runoff responds rapidly to rainfall and the translation effect is dominant. With increasing drainage area and lower basin slopes, temporary storage becomes more important to give a decreasing discharge per unit area and a smaller value of n in equation (3.3). Thus, the shape of the channel and network geometry influence stream behaviour in the short term. Attempts to evaluate the effects of the latter on hydrograph characteristics have suffered from the lack of an adequate theoretical basis which would help to separate individual influences.

Natural river flow is highly variable in time and space. Partly because of the increasing availability of flow records, discharge has become a primary independent variable in geomorphological approaches to the description and analysis of river channel form (Figures 1.2B, 4.7). The importance of discharge can perhaps be best appreciated from the dominant-discharge concept and the hydraulic geometry approach pioneered by Leopold and Maddock (1953). However, discharge is a summary variable which does not express directly the forces involved in shaping channels. To that extent the role of discharge has possibly been overemphasized. A closer link is needed between channel form adjustment and the mechanical work performed by streams, for which background theory is available.

Mechanics of flow

Water flowing in an open channel is subject to two principal forces: gravity, which acts in the downslope direction to move water at an acceleration of $g\sin\beta$ where g is gravitational acceleration (= 9.81 m s^{-2}) and β the angle of slope; and friction, which opposes downslope motion. The relationship between these two forces ultimately determines the ability of flowing water to erode and transport debris.

Open channel flow can be classified into various types based on four criteria (Table 3.1). Simple mathematical models can be constructed only if the flow is

Table 3.1 Types of flow in open channels

Type of flow	Criterion
Uniform/Non-uniform (varied)	Velocity is constant/variable with position
Steady/Unsteady	Velocity is constant/variable with time
Laminar/Turbulent	Reynolds' Number ($R_e = vR\rho/\mu$) is $< 500/ > 2\,500$, with a transitional type when $500 < R_e < 2\,500$
Tranquil/Rapid	Froude Number ($F = v/\sqrt{gd}$) is $< 1/> 1$, with critical flow when $F = 1$

Symbols: v, velocity; R(= w.d/2d + w), hydraulic radius; w, width; d, depth of flow; ρ, density; μ, viscosity; g, gravity constant

assumed to be uniform and steady but flow in natural rivers is characteristically non-uniform and unsteady. The important distinction lies between laminar and turbulent flow.

Water is a viscous fluid that cannot resist stress, however small. In *laminar flow* each fluid element moves along a specific path with uniform velocity and no significant mixing between adjacent layers. A very thin layer of fluid in contact with the boundary is slowed so completely that it has no forward velocity but resistance to motion along internal boundaries is less than at the bed and each successive layer of fluid away from the bed can slip past the one below to give a velocity profile which is parabolic in shape (Figure 3.3A). In this way is shear stress,

$$\tau = \mu \frac{dv}{dy} \tag{3.4}$$

distributed throughout the flow, where dv/dy is the velocity gradient at depth y.

When velocity or depth exceeds a critical value, laminar flow becomes unstable and the parallel streamlines are destroyed. In *turbulent flow* the fluid elements follow irregular paths and mixing is no longer confined to molecular interactions between adjacent layers but involves the transfer of momentum by large-scale eddies. Accordingly the equation for shear stress must be modified to include an eddy viscosity (η) term,

$$\tau = (\mu + \eta) \frac{dv}{dy} \tag{3.5}$$

Because $\eta \gg \mu$ turbulent flow exerts larger shear stresses than does laminar flow for the same velocity gradient (dv/dy). Also, velocity tends to be more evenly distributed with depth because of the larger-scale mixing which slows faster bodies of water up the profile and speeds up slower ones below (Figure 3.3A).

The mathematical analysis of laminar flow is well advanced but this type of flow rarely occurs in nature except as a very thin layer (the laminar sublayer) close to the channel boundary or as shallow overland flow (Emmett, 1970). At first sight the equation for the velocity profile (3.5) appears to offer a means of evaluating in theory the shear stress acting at the bed (τ_0), an important force in grain movement. However, turbulent flow is so complex and irregular that a suitable physical model has yet to be formulated and analysis relies heavily on experimentation. Representative values of eddy viscosity and the velocity gradient at the bed are difficult if not impossible to obtain. Consequently the possibility of applying relatively simple models to the problem of fluid transport is denied.

Velocity and resistance

Velocity is a vector quantity having both magnitude and direction. It is one of the most sensitive and variable properties because of its dependence on most of the factors which characterize open channel flow.

Velocity varies in four dimensions (Figure 3.3):

(i) With distance from the stream bed – Aside from its variation with the type of flow (laminar or turbulent), the shape of the velocity profile is strongly influenced by the size of roughness elements on the stream bed and the depth of flow (d). If the former is expressed in terms of bed material size (D), the two variables can be incorporated in a single index, the relative roughness ratio d/D. For a given depth of flow, the larger the roughness elements the steeper is the velocity gradient toward the bed. Since stream beds generally contain a range of grain sizes, the problem is to select a grain diameter which best expresses this component of resistance.

(ii) Across the stream – Velocity increases toward the centre of a stream as the frictional effects of the channel banks decline but the degree of symmetry in cross-channel velocity can be highly variable, changing with the shape and alignment of the channel. In particular the velocity distribution in channel bends is characteristically asymmetric with the main current moving toward the outer bank. The close relationship between velocity distribution, cross-sectional shape and erosive tendency is emphasized by the basic distinction between wide, shallow channels where the velocity gradient is steepest and the boundary shear stress therefore greatest (equation 3.5) against the bed, and narrow, deep sections where the velocity gradient is steepest against the banks, producing a greater tendency for bank erosion (Figure 3.3B).

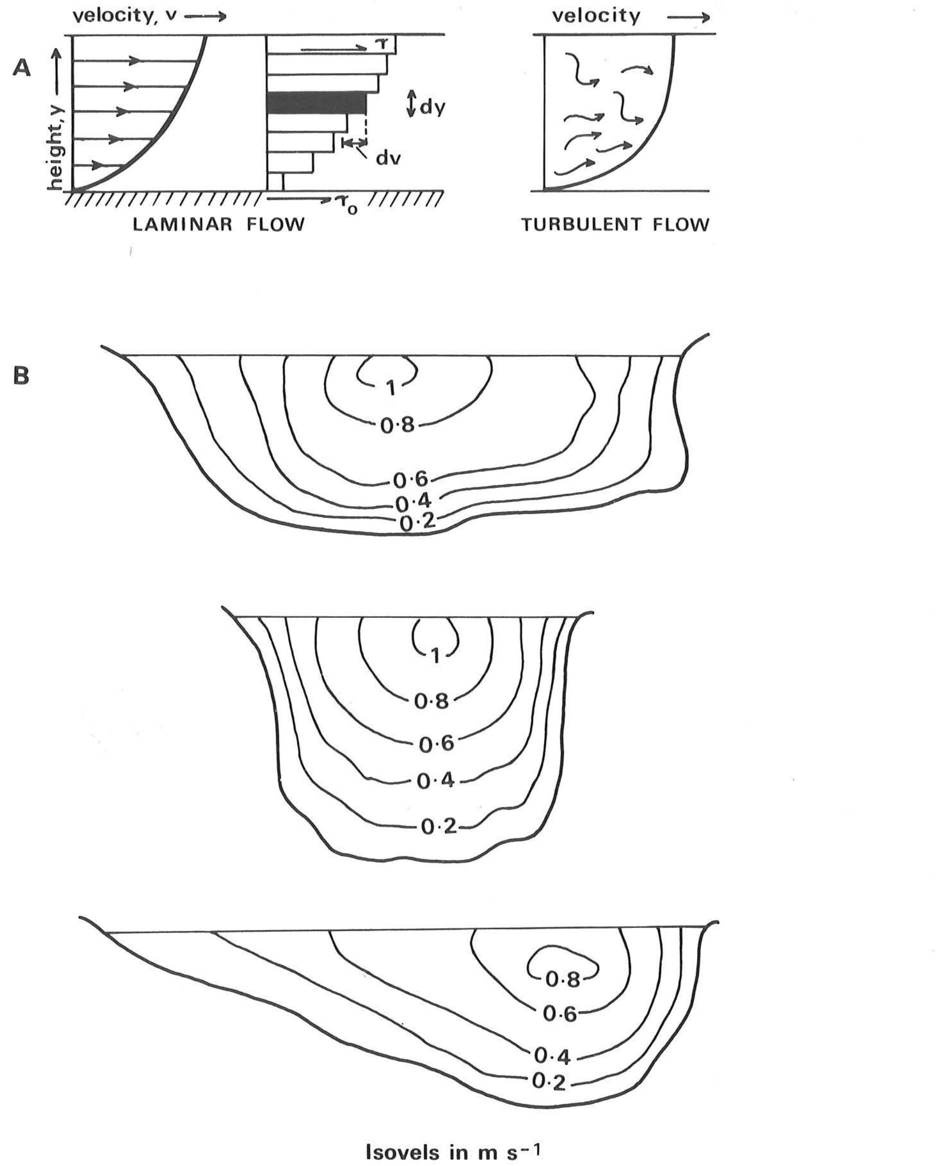
(iii) Downstream – In addition to local fluctuations, change in velocity at the longitudinal scale has been a major interest of geomorphologists concerned in particular with the development of an equilibrium stream profile. Despite a declining slope along most rivers, velocity tends to remain constant or increase slightly (Carlston, 1969) as the channel becomes hydraulically more efficient and resistance decreases in the downstream direction. Variations in the rate of change of velocity do occur both along and between rivers, since velocity is merely one variable that can be adjusted to accommodate the downstream increase in discharge.

(iv) With time – Over time periods measured in seconds point velocities may reach values of 60–70 per cent or more of the time average velocity because of the inherent variability of turbulent flow, thus making it difficult to define the initiation of particle motion in terms of velocity. At the larger time scale of days, weeks or months, velocity responds to fluctuations in discharge. The increase in depth with discharge tends to drown out roughness elements in the bed and thereby produce an increase in velocity. However, the effect is not uniform and the exponent m (the rate of change of velocity) in

$$v = kQ^m \tag{3.6}$$

can vary considerably from section to section (Park, 1977).

Velocity is thus a highly variable quantity in time and space. The character of that variation is important since velocity influences the processes of erosion, transportation and deposition (Figure 3.5D). Velocity is usually measured by current



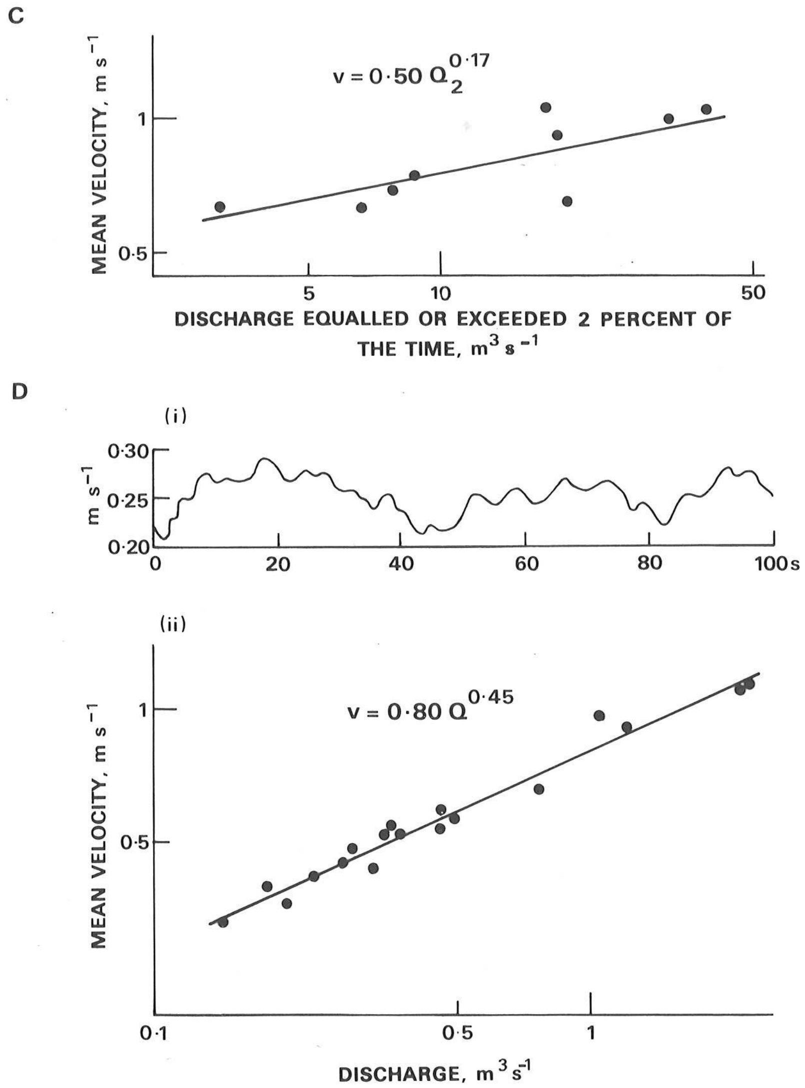


Fig 3.3 Variations in streamflow velocity:
A. With depth – typical velocity profiles for laminar and turbulent flow are shown.
B. At natural channel cross-sections.
C. Downstream – relationship of velocity to discharge, Brandywine Creek (after Wolman, 1955).
D. With time – (i) velocity fluctuations at a point over a short time period, (ii) at-a-station changes in velocity with discharge measured over two years, River Bollin.

meter at selected points in the flow cross-section and expressed as an average value. However, mean velocity at a cross-section is not the most relevant measure for defining the initiation of erosion but it remains the most widely used parameter, partly because of measurement difficulties close to the stream bed.

Velocity is strongly related to *flow resistance*, one of the most important elements in the interaction between the fluid flow and the channel boundary. Several resistance equations have been developed (Table 3.2), of which the Darcy–Weisbach is recommended for its dimensional correctness and sounder theoretical basis (Task Force, 1963). All of the equations assume that resistance approximates that of a steady, uniform flow but in natural channels with erodible boundaries the resistance problem is much more involved.

Table 3.2 Flow resistance equations

Chezy equation (1769)	$v = C\sqrt{Rs}$	
Manning equation (1889)	$v = \frac{KR^{\frac{2}{3}} s^{\frac{1}{2}}}{n}$	where K = 1 (SI units), K = 1.49 (imperial units)
Darcy–Weisbach equation	$ff = \frac{8gRs}{v^2}$	

Symbols: C, n, ff are the respective resistance coefficients;
v, mean velocity; R, hydraulic radius;
s, slope of the energy gradient; g, gravity constant

Resistance in a flow carrying sediment comprises several components: grain or surface roughness, form roughness, channel irregularities, and suspended material in the flow. *Grain roughness* is a function of relative roughness (d/D or R/D) and in rough channels with fixed boundaries is often expressed in the form

$$\frac{1}{\sqrt{ff}} = c \log \left(a \frac{R}{D_x} \right) \qquad (3.7)$$

where c and a are constants and D_x is a measure of the size of roughness elements (Figure 3.4A). While the roughness height of uniform material is simply taken as the grain diameter, the problem with non-uniform material where the effective size exceeds the mean is to choose a representative grain diameter. No one diameter has been generally accepted as suitable for this role, although D_{84} is commonly used as a roughness height. D_{90}, the diameter which equals or exceeds 90 per cent of the material, has been suggested for British gravel rivers (Charlton *et al.*, 1978). A single figure may not be representative of grain roughness at cross-sections where the grain size distribution varies markedly across the bed; nor are the effects of bank friction explicitly included in equation (3.7) (Hey, 1979).

Grain roughness is probably the dominant component of resistance where stream beds consist of gravel (2–64 mm) or cobbles (64–256 mm). Equation (3.7) predicts that, as depth increases with discharge at a cross-section, the effect of grain roughness is drowned out and flow resistance decreases, although possibly at a declining rate with higher discharge (Figure 3.4B). Consequently velocity may also tend to change more slowly at higher flows, producing non-linearities in hydraulic geometry (Richards, 1973, 1977; Figure 4.7B).

Form roughness stems from features developed in the bed material and often exceeds grain roughness in importance. It presents a particular problem in that, once grains are set in motion, the shape of the bed can be modified to give a variable form roughness dependent on flow conditions. In sand-bed streams where the bed is most readily moulded into different shapes, a sequence of bed forms correlated with increasing velocity has been defined (Figure 3.4C), each form offering different levels of resistance (Simons and Richardson, 1966). However, understanding of the relationship between bed form geometry and hydraulic roughness remains incomplete and attempts to predict the occurrence of different bed forms in terms of selected flow, fluid and sediment properties have met with only qualified success (Lawson and O'Neill, 1975). Nevertheless, changes in bed configuration represent an important self-regulating mechanism available to streams at the interface of the fluid flow with the erodible bed. As discharge and sediment load increase with the passage of a flood wave, a transition from ripples to dunes

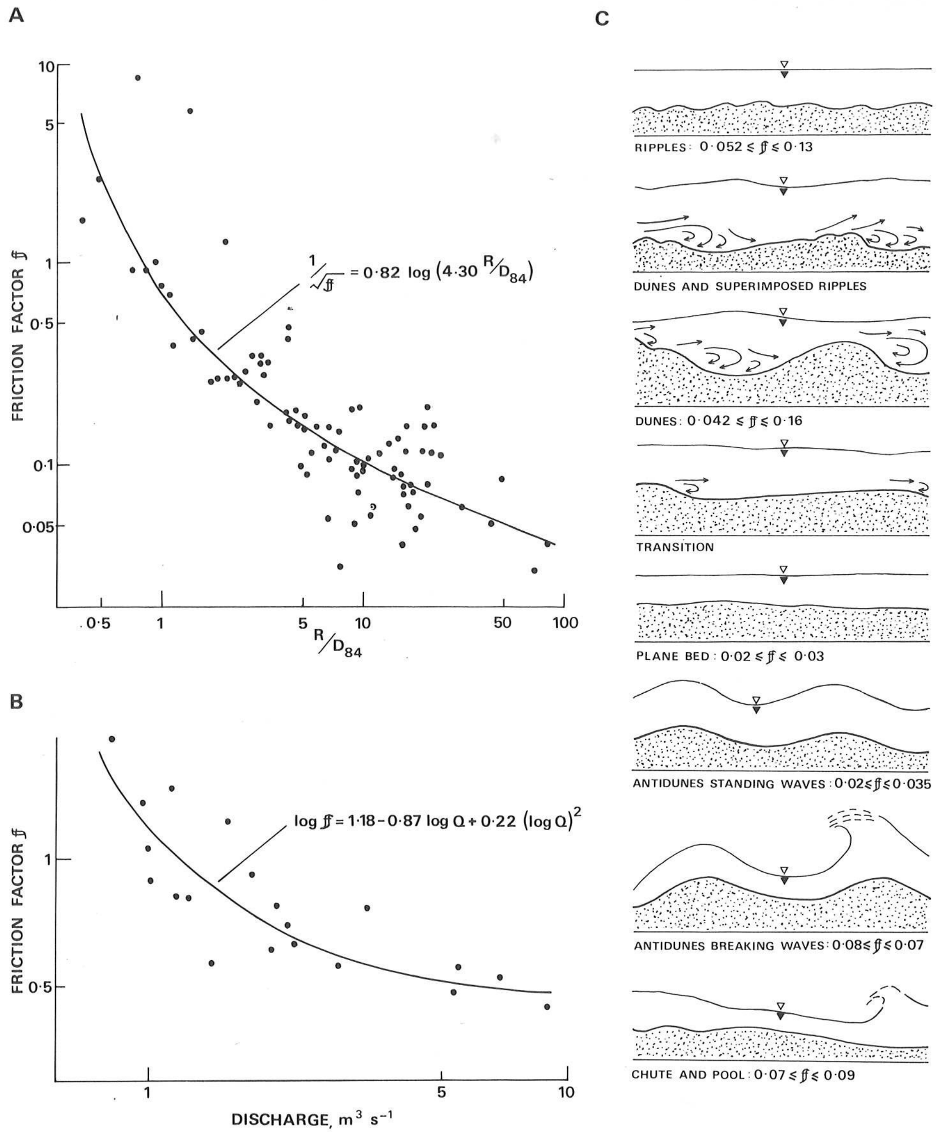


Fig 3.4 A. Relationship between friction factor and relative roughness (data of Leopold and Wolman, 1957; Limerinos, 1970; Charlton et al., 1978; Hey, 1979).
B. Relationship between friction factor and discharge at a sandy-bed cross-section, River Bollin.
C. The sequence of bed forms related to increasing flow intensity, with corresponding values of the Darcy-Weisbach friction factor in flume experiments (after Simons and Richardson, 1966).

may so increase the flow resistance as to offset the improved hydraulic efficiency associated with a larger depth, and thereby slow the rate of change of velocity (Richards, 1973, 1977).

Dunes are rarely found in gravel-bed streams where bed mobility is more restricted. There, the roughness due to bars becomes increasingly important at lower flows with grain roughness dominant at higher ones (Parker and Peterson, 1980). However, in streams with very coarse beds the form drag exerted by individual particles can be considerable. Bathurst (1978) distinguished two roughness regimes based on whether relative roughness (d/D) is less than or greater than about 3; form roughness is dominant in the first and grain roughness in the second. Resistance to flow under such conditions remains rather poorly defined.

The remaining components of resistance due to *channel irregularity* and *suspended matter* have received comparatively little attention. Natural channels are characteristically irregular and considerable energy loss can occur because of local bank irregularities and changes in channel alignment. Indeed channel curvature may introduce an energy loss greater than that associated with grain roughness (Leopold *et al.*, 1960).

Material suspended in the flow tends to damp down turbulence and thereby reduce resistance. For a fixed dune bed, Vanoni and Nomicos (1960) showed experimentally that suspended sediment concentrations of 3.64 and 8.08 kg m^{-3} decreased friction by 5 and 28 per cent respectively relative to clear-water flow. However, data are rather sparse and the general opinion is that suspended material has a relatively small influence on flow resistance (Raudkivi, 1976).

All treatments of the resistance problem tend to be oversimplified. Neither the flow, form nor roughness of individual cross-sections is uniform. Resistance is invariably computed from one of the equations (Table 3.2) using hydraulic data rather than by direct measurement, so that computed values include the effects of all types of roughness. Attempts to separate total resistance into its component parts (e.g. Einstein and Barbarossa, 1952) have achieved only qualified success. Much of the empirical work on resistance has been carried out under controlled conditions in laboratory flumes and it is not entirely clear how applicable are the results to the field situation where energy loss can be highly localized. In short, knowledge of the resistance mechanism remains far from complete, particularly in natural streams. And yet flow resistance is a primary concern through its link with sediment transport and the way in which a stream consumes its energy. Indeed Davies and Sutherland (1980) have proposed that stream behaviour and channel development are governed by a principle of maximum resistance in which boundary deformation continues until flow resistance has attained local maxima, a hypothesis that remains to be tested.

Stream energy

Energy is a quantity expressed as

force (mass x acceleration) x distance through which the force acts

and having the dimensions of $ML^2 T^{-2}$. In the fluvial system three types of energy are relevant – potential, kinetic and thermal (heat) – only the first two of which can perform mechanical work. That work takes various forms:

(i) work done against viscous shear and turbulence (internal friction);
(ii) work done against friction at the channel boundary;
(iii) work done in eroding the channel boundary;
(iv) work done in transporting the sediment load.

Since energy must first be used to maintain the flow against internal and boundary friction ((i) and (ii)), a critical energy level must be reached before a stream can perform erosional and transportational work. The concept of an erosion threshold is fundamental and needs to be incorporated in sediment transport equations (e.g. Bagnold, 1977).

Water with a mass m entering a river at a height h above a given datum (base-level or the next tributary junction) has a potential energy

$$PE = mgh \tag{3.8}$$

As water moves downslope that potential or position energy is gradually converted into kinetic energy

$$KE = \tfrac{1}{2} mv^2 \tag{3.9}$$

In conservative systems the principle of energy conservation

$$PE + KE = \text{constant} \tag{3.10}$$

applies and relatively simple models can be developed for describing system behaviour. Between any two points, a loss in potential energy is matched by an equivalent gain in kinetic energy. However, rivers are non-conservative systems. Friction causes much of the available mechanical energy to be dissipated in the form of heat which can perform no mechanical work. Consequently the well established principles of mechanics cannot readily be used to solve flow problems without making far-reaching assumptions.

Nevertheless, it is relevant to ask how potential energy could be distributed in the fluvial system since different distributions may be associated with different network topologies (Knighton, 1980a). A precise formulation has yet to be achieved but could provide a basis for modelling stream behaviour and establishing network–channel links.

The importance of energy distribution has been recognized in the development of theoretical models concerned with the search for a general principle governing stream behaviour and channel form adjustment. Arguing by analogy between fluvial and thermodynamic systems, Langbein and Leopold (1964) postulated that the most probable or modal state represents a compromise between a uniform distribution of energy expenditure and minimum total work. Yang (1971a) argued that, in moving toward a state of dynamic equilibrium, a natural stream chooses its course in order to minimize the time rate of potential energy expenditure, manifest in the development of meanders (Yang, 1971b) and riffle–pool sequences (Yang, 1971c).

However, direct considerations of energy expenditure in natural streams are very few. Nor has the amount of work required to move from one channel state to another been directly calculated despite its relevance to models of channel form adjustment. There is a need to relate more closely the activity of fluvial processes and the forms they develop to the physical concept of work. Bagnold (1966, 1977) has led the way in relating sediment transport rate to available stream power, where the power (or rate of doing work) per unit length of stream is

$$\Omega = \gamma Qs \tag{3.11}$$

where γ (= ρg) is the specific weight of water, Q is discharge and s is slope. Power so defined is the rate of potential energy expenditure per unit length of channel.

Thresholds of erosion

The movement of particles depends on their physical properties, notably size, shape and density. Grain size has a direct influence on mobility and a typical classification is shown in Table 3.3. A base distinction exists between non-cohesive and cohesive (including solid rock) materials. In cohesive sediments, which generally consist of particles in the silt–clay range, resistance to erosion depends more on the strength of cohesive bonds between particles than on the physical properties of particles themselves, making the erosion problem much more complex.

Table 3.3 Grain-size classification

Class name	Size range mm	phi units
Boulders	⩾256	⩽-9
Cobbles	64–256	−6 to −9
Gravel	2–64	−1 to −6
Sand	0.064–2	4 to −1
Silt	0.004–0.064	8 to 4
Clay	⩽0.004	⩾8

Bed erosion

The characteristics of stream bed material depend on the initial supply conditions and the subsequent action of such processes as sorting and abrasion. Most stream beds consist of cohesionless grains. As the flow over a surface of loose grains gradually increases, a condition is reached when the forces tending to move a particle are in balance with those resisting motion. The problem is to define this threshold state. Three related approaches have been used, initial movement being specified in terms of either a critical shear stress (τ_{cr}), a critical velocity (v_{cr}) or the lift force.

A reasonable estimate of the boundary shear stress (τ_o) exerted by the fluid on the bed can be obtained from

$$\tau_o = \gamma Rs \tag{3.12}$$

where γ is the specific weight of water, R is hydraulic radius and s is slope. Derived from a consideration of the balance of forces in a steady non-accelerating flow (see Leopold *et al.*, 1964, p. 156–7), this quantity is a spatial average which does not necessarily provide a good estimate of bed shear at a point. Recognizing this limitation, the critical shear stress (τ_{cr}) can then be defined by equating the two sets of forces involved:

applied forces – fluid forces and the downslope component of the particle's submerged weight

resisting forces – the component of the particle's submerged weight acting normal to the bed and any constraining forces due to neighbouring grains.

For spherical grains of diameter D on a flat bed, equating the moments of forces acting about a downstream contact point (A in Figure 3.5A) gives

$$\tau_{cr} = \eta\, g\, (\rho_s - \rho)\, \frac{\pi}{6}\, D \tan\phi \tag{3.13a}$$

This elementary deterministic model predicts that the shear stress needed to initiate movement increases with particle size, grain shape and the degree of packing (η) also being influential.

Equation (3.13a) has been expressed in many ways, most notably by Shields (1936) who recognized that critical shear stress depends not only on particle size but also on bed roughness. The resultant plot (Figure 3.5C) relates a dimensionless critical shear stress (θ) to a particle Reynolds number ($\propto D/\delta_o$, where δ_o is the thickness of the laminar sublayer) which defines the bed roughness condition. The plot not only separates zones of no motion and motion but reveals that the relationship between threshold stress and particle size is not as straightforward as a simple resolution of forces would at first suggest. On hydraulically rough beds (the common condition in natural streams), θ rapidly attains a constant value of 0.06 to give thereafter

$$\tau_{cr} = 0.06\, g\, (\rho_s - \rho)\, D \tag{3.13b}$$

However, at lower stresses the thickness of the laminar sublayer increases and, for particles less than about 0.7 mm, the relationship becomes inverse so that the threshold stress needed for entrainment must increase as particles get smaller. Particles in that size range are submerged in the laminar sublayer and therefore not subject to the greater stresses in the overlying turbulent layer.

An alternative approach defines the critical condition in terms of velocity rather than shear stress, but the same basic trends are revealed (Figure 3.5D). Hjulström's empirical curve predicts that medium sand (0.25–0.5 mm) is the most easily eroded fraction, higher velocities being required to set both coarser and finer grains in motion. Mean velocity is not the most relevant parameter in this context but the problem is to define and measure a bottom velocity representative of the threshold condition.

These approaches have important limitations. They fail to cope adequately with the variability of either flow conditions near the stream bed (Figure 3.3D(i)) or bed material characteristics. Short-term pulsations in the flow can give rise to instantaneous stresses of at least three times the average, so that particles may be entrained at stresses much lower than predicted. Sediment entrainment is a function not only of the average shear stress on the bed but also of the intensity of turbulence above it, which can exert an impulse force on grains. Since eddy size and hence the energy available for moving grains are related to the size of the system, the size of the channel can influence the entrainment process (Raudkivi, 1976). Natural bed material is neither spherical nor of uniform size. Larger particles may shield smaller ones from direct impact so that the latter fail to move until higher stresses are attained. Characteristics other than size influence particle mobility, notably the degree of grain exposure (Fenton and Abbott, 1977), bed relief and sediment fabric (Laronne and Carson, 1976). Thus for a given grain size > 8 mm, the Shields' threshold criterion may vary by nearly an order of magnitude depending on whether the bed is loosely or tightly packed (Church, 1978). Consequently empirical and theoretical attempts to define the threshold for grain entrainment

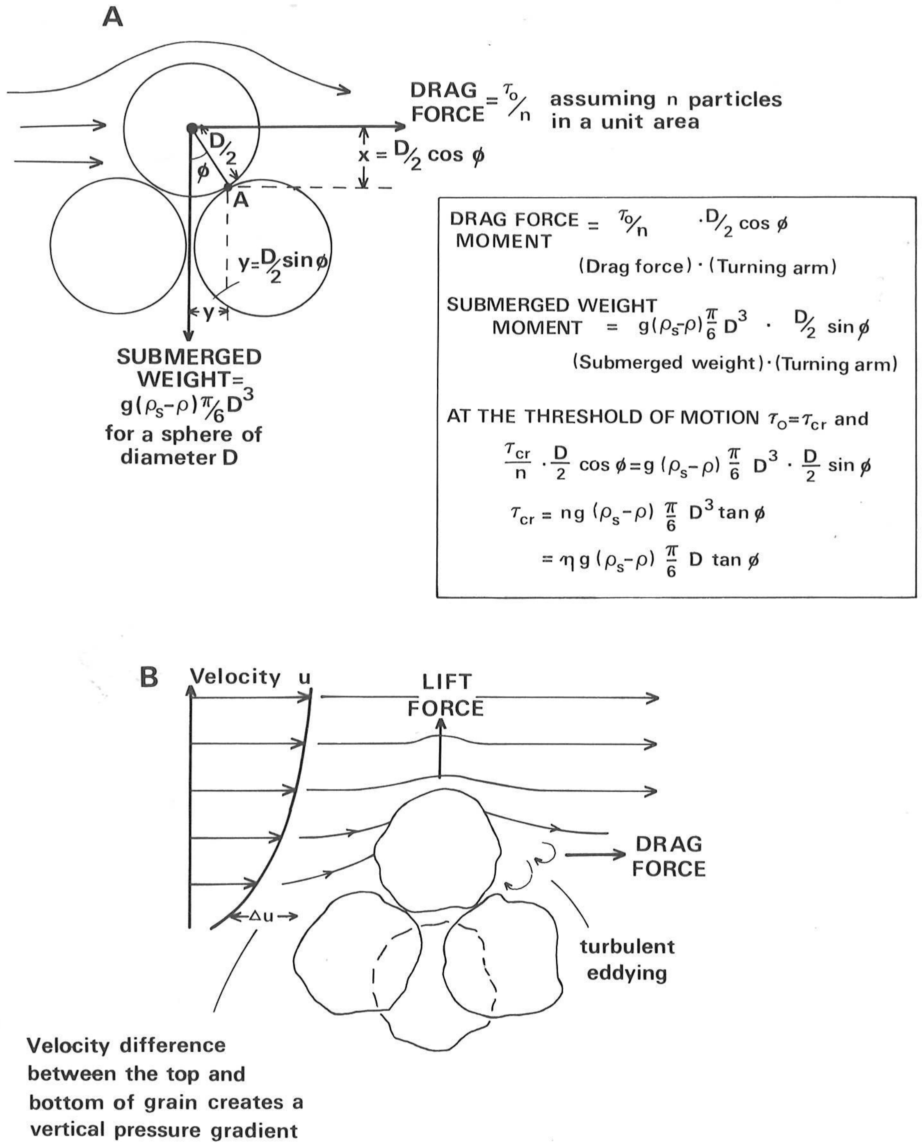
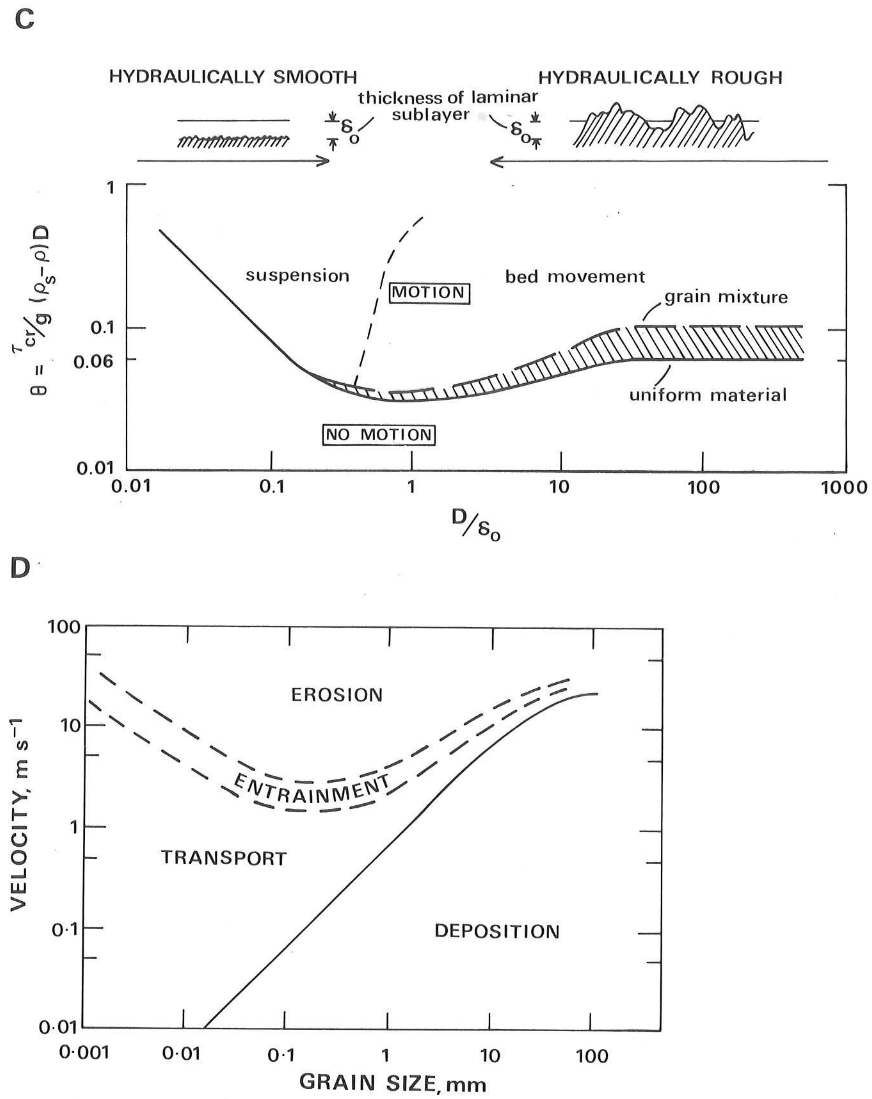


Fig 3.5 A. The critical shear stress (τ_{cr}) defined for a grain resting on a horizontal bed, where ρ is fluid density, ρ_s sediment density, g gravity constant, D grain diameter, and η = nD^2 a measure of grain packing.
B. Lift and drag forces acting on a submerged particle.
C. The Shields' entrainment function.
D. Erosion and deposition criteria defined in terms of threshold velocities (after Hjulström, 1935).

can at best predict an average state. The inherent variability of natural flow and bed material conditions gives a statistical character to the problem and this has led to the development of probabilistic models of particle motion (e.g. Hung and Shen, 1976).



In addition to the drag forces acting roughly parallel to the bed, there is a force (the *lift force*) normal to the bed which may be able to entrain particles irrespective of the magnitude of the drag forces (Figure 3.5B). This force arises in at least two ways:

(i) the difference in flow velocity between the top and bottom of a grain sets up a pressure gradient which tends to move the particle vertically upwards;
(ii) turbulent eddying may produce local velocity components which act directly upwards close to the bed.

The lift force decreases rapidly in magnitude away from the bed as the velocity and pressure gradients become less. There is no generally accepted theory applicable to the lift force and little experimental data available for its estimation. The relative contributions of the lift and drag components are also difficult to determine because of their interrelationship but the lift force may be a crucial mechanism for initiating grain movement, at least in sand-bed streams. The combination of lift and drag provides one explanation for the commonly-observed process of saltation (Figure 3.8A) in which particles of sand size move along the bed in a series of hops.

Much of the work on the threshold condition has been carried out in laboratory flumes with relatively fine material. Helley's (1969) work on the movement of large particles (> 150 mm in diameter) is one of the few field studies and, perhaps surprisingly in view of previous comments, the bed velocities required for entrainment agreed quite closely with those predicted by a deterministic model which included the effects of particle size, shape and orientation. Threshold velocities existed for at least 5 per cent of the time in this coarse bed stream and, although the sample was small, the results support Hjulström's curve (Figure 3.5D). However, it appears that in general both the Hjulström and Shields curves require adjustment when predicting the erosion threshold in natural streams with bed material coarser than sand (Novak, 1973; Baker and Ritter, 1975).

The erosion of cohesive sediment presents a different and more complex problem. The forces resisting motion include not only those associated with particle weight but also the electro-chemical forces which bind the material. Such forces cannot be uniquely expressed as a function of particle size or similar variables. The limited data available suggest that there is no critical shear stress for the erosion of cohesive material in the sense that one exists for cohesionless grains. Erosion takes place in aggregates rather than particle by particle.

Similar remarks apply to bedrock channels as regards the erosion threshold. Several processes have been identified in general terms:

(i) corrosion, or the chemical action of water;

(ii) corrasion, or the mechanical (hydraulic and abrasive) action of water which can be effective when the flow is armed with particles, leading to surface abrasion and pothole development;

(iii) cavitation, a process associated with the effects of shock waves generated through the collapse of vapour pockets in a flow with marked pressure changes.

Cavitation can cause severe erosion (Brown, 1963) but is of limited occurrence because of the high velocities required. These processes remain little more than descriptive terms. However, Foley (1980) has attempted to model abrasion using engineering sandblast theory, distinguishing between low- and high-angle impacts which respectively cause abrasion through cutting and fracturing.

The concept of a critical shear stress or velocity necessary for erosion is appealing, at least in theory. Theory and observation indicate that material in the size range 0.25–1 mm is the most susceptible to movement and, other conditions being equal, will be the first to be entrained. However, the many factors involved are difficult to accommodate theoretically. The heterogeneity of natural stream beds and the variability of flow conditions in the bed region limit the applicability of deterministic models and results obtained from laboratory studies, but field tests remain very difficult to make. A further dimension to the problem is provided by the cohesivity of some stream beds where threshold criteria appear to be less relevant.

Bank erosion

The bank material of natural channels is even more variable than the bed material but does tend to become finer and more uniform downstream with combinations of sands, silts and clays dominant, especially where the flood-plain is well developed. Most channel banks possess some degree of cohesion because of finer material, so that the analysis of bank erosion is not a simple extension of the non-cohesive bed case with a downslope gravity component added. A further complication is provided by the effects of vegetation whose root system can reinforce bank material and thereby increase resistance to erosion. Smith (1976) found that bank sediment with a root volume of 16–18 per cent and a 5 cm root mat afforded 20,000 times more protection from erosion than comparable sediment without vegetation. This effect may be more limited in large rivers with high banks where root systems do not protect lower bank areas.

Bank erosion is one of the principal means of sediment supply to streams. Klimek (1974) estimated that about 2000 m^3 of sediment were supplied each year from a 100 m section of the River Wisloka in Poland. Adjustments to the form and course of river channels and the development of flood-plains depend on bank erosion, which also threatens man-made structures and destroys valuable agricultural land. However, despite its overall importance, relatively few field or laboratory studies have been undertaken.

The main processes involved can be grouped under four headings: direct action of water, slumping, rotational slipping and frost action. The shearing of bank material by *hydraulic action* at high discharges is a most effective process, especially on non-cohesive banks and against bank projections. Large-scale eddying induced by bank irregularities can enlarge existing embayments and increase the amplitude of projections which become more susceptible to subsequent attack. In cohesive banks the exploitation of pre-existing cracks can lead to the removal of joint-bounded blocks which may accumulate at the bank foot and afford temporary protection. These several effects underline the importance of shear stress distribution and local turbulence characteristics in erosion by hydraulic action. In many instances it is the dominant process (Knighton, 1973; Hooke, 1979) and Simons *et al.* (1979) estimated that flow forces were at least 6 times more effective than any other process in their study of the Connecticut River. Those forces can be subdivided into:

(i) those which act near the surface of the flow, such as water waves induced by wind or passing boats; and

(ii) those which act near the base of banks and lead to undercutting.

Significantly, velocity and boundary shear stress appear to be at a maximum in the lower bank region even when the flow is near bankfull (Bathurst *et al.*, 1979; Simons *et al.*, 1979).

The collapse or *slumping* of large blocks of material is more closely related to soil moisture than to flow conditions, although oscillations in river stage can influence the degree of bank wetting. Cohesive banks are particularly susceptible to seepage forces and piping mechanisms that may so lower the internal resistance of the material as to induce failure. Wet bank slumping sometimes takes place after the main flow has receded when the bank is thoroughly wetted (Twidale, 1964) and can be a major contributor to bank retreat (e.g. Klimek, 1974). The process is influenced by bank stratigraphy, for example where cohesive materials overlie non-cohesive ones, a relatively common condition in rivers flowing through alluvial deposits. In the River Wisloka, water percolating through the basal region loosens the sandy-gravel and liquifies the lower parts of the overlying silty-clay, thereby inducing slumping over the entire bank face (Klimek, 1974). Analysing the stability of composite banks, Thorne and Tovey (1981) argued that, whereas the lower bank is eroded by hydraulic action, the upper bank is less affected by flow forces but fails

Table 3.4 Factors influencing bank erosion

Factor	Relevant characteristics
Flow properties	Magnitude, frequency and variability of stream discharge Magnitude and distribution of velocity and shear stress Degree of turbulence
Bank material composition	Size, gradation, cohesivity and stratification of bank sediments
Climate	Amount, intensity and duration of rainfall Frequency and duration of freezing
Subsurface conditions	Seepage forces, piping Soil moisture levels
Channel geometry	Width and depth of channel Height and angle of bank Bend curvature
Biology	Type, density and root system of vegetation Animal burrows
Man-induced factors	Urbanization, land drainage, reservoir development, boating

because of undercutting which produces different types of cantilever action in the cohesive material. The slumped blocks may break on impact or remain intact to await removal by subsequent flows, in the meanwhile protecting the underlying gravel from further erosion. This pseudo-cyclic process (undercutting, upper bank failure, and removal of failed blocks) has been observed elsewhere (Stanley *et al.*, 1966) and emphasizes the interrelationship of different processes.

The remaining processes are probably less significant. *Rotational slipping* produces multi-stepped bank profiles through repeated failure along surfaces roughly concave to the slope, a process whose effectiveness is again strongly related to moisture levels in the bank. Where and when climatic conditions allow, the growth of ice crystals or ice wedges may affect bank retreat (Walker and Arnborg, 1966). However, *frost action* is more important as a preconditioning than an erosive process in that it widens pre-existing cracks and disaggregates surface material to leave the bank more susceptible to subsequent attack. This effect again emphasizes that, although hydraulic action and slumping may be dominant, the several processes are not mutually exclusive but frequently act in combination.

The amount and periodicity of river bank erosion are highly variable because of the large number of factors involved (Table 3.4), and average rates (Table 3.5) tend to mask that variability. Little erosion is likely to occur without high discharges but similar flows need not be equally effective because they may not be attacking against a bank in the same condition. Consequently correlations between flow volume and amount of erosion tend to be rather weak. Wolman (1959) found that a large summer flood attacking dry banks produced little erosion but that lesser winter flows acting against thoroughly wetted banks caused considerable bank retreat. Bank wetting and frost action are important preconditioning processes which reduce the strength of bank material and make it more susceptible to erosion. Therefore, multi-peaked flows may be more effective than single flows of comparable or greater magnitude because of the increased incidence of bank wetting (Knighton, 1973). The main point is that the amount of bank erosion is not solely a function of discharge magnitude or even flow conditions, so that a threshold flow cannot reasonably be defined, although at some of her sites Hooke

Table 3.5 Measured rates of bank erosion

River and location	Drainage area, km²	Average rate of bank retreat, m yr⁻¹	Period of measurement	Source
Axe, Devon	288	0.15–0.46	1974–76	Hooke (1980)
Bollin-Dean, Cheshire	12–120	0–0.9	1967–69	Knighton (1973)
Cound, Shropshire	100	0.64	1972–74	Hughes (1977)
Crawfordsburn, N. Ireland	3	0–0.05	1966–68	Hill (1973)
Exe, Devon	620	0.62–1.18	1974–76	Hooke (1980)
Mississippi, Louisiana	–	4.5	1945–62	Stanley *et al.* (1966)
Torrens, S. Australia	78	0.58	1960–63	Twidale (1964)
Watts Branch, Maryland	10	0.5–0.6	1955–57	Wolman (1959)
Wisloka, Poland	–	8–11	1970–72	Klimek (1974)

(1979) found little or no erosion at discharges with a frequency greater than 5 per cent.

Sites experiencing the same flow and meteorological conditions can show considerable variation in the amount of bank erosion (Knighton, 1973; Hooke, 1979). Flow volume and bank moisture levels may provide the conditions necessary for erosion but by themselves they are not sufficient. The local site characteristics which appear to have a major influence on the spatial distribution of erosion are bank material composition, the degree of flow asymmetry and channel geometry. Coarser, sandy materials are more liable to erosion than are those with a high silt-clay content. In composite banks, stability is governed by the strength of the weakest material since its removal will eventually produce failure in the rest of the bank (Krinitzsky, 1965; Thorne and Tovey, 1981). Flow asymmetry was necessary for erosion along the Bollin-Dean as only then were velocities sufficiently high close to the bank face (Knighton, 1973). Not only may the distribution of velocity or shear stress change with discharge but also between reaches of different curvature to give locally varied rates of bank retreat (Hickin and Nanson, 1975).

With so many factors involved (Table 3.4), it is hardly surprising that their individual effects cannot be separated or an erosional threshold defined. Field work has identified some of the relevant variables but measurement periods tend to be rather short (Table 3.5). The need remains for detailed research into the mechanics of bank erosion with increasing emphasis on the distribution of forces against river banks and the changeable quality of bank resistance.

Sediment transport

The transport of material from the land surface to the sea can be rationalized in terms of three process regimes (Statham, 1977):

(i) a *weathering regime* which includes those processes involved in the physical and chemical breakdown of rocks;

(ii) a *slope regime* in which the products of weathering are moved down the gravity gradient in mass movements and by slope wash processes; and
(iii) a set of *fluid-transfer regimes*, water, air and ice, of which the first is by far the most important.

All material entering a river system must cross the boundary between the slope and fluvial regimes provided by the channel banks and the channel head.

An important distinction is between supply-limited and capacity-limited transport. Much of the material supplied to streams is so fine that, provided it can be carried in suspension, almost any flow will transport it. Although an upper limit must exist in theory, the transport of this fine fraction is largely controlled by the rate of supply rather than the transport capacity of the flow. In contrast, the transport of coarser material (> 0.064 mm) is capacity-limited and therefore intermittent, otherwise channels would show a greater tendency to have beds of solid rock rather than cohesionless grains. The intermittency of bed-material transport and the possibility of temporary deposition mean that the residence times of material moving through even small drainage basins are likely to be large, of the order of 100s or 1000s of years.

Figure 3.6 illustrates the main elements in the movement of material through the fluvial system. The load carried by natural streams can be separated into three components:

(i) the *dissolved load*, consisting of material transported in solution;
(ii) the *wash load*, comprising particles finer (< 0.064 mm) than those usually found in the bed and moving readily in suspension; and
(iii) the *bed-material load*, including all sizes of material (> 0.064 mm) found in appreciable quantity in the bed.

The bed-material load may be transported as *bed load*, when particles move by rolling, sliding or saltation at velocities less than those of the surrounding flow, or as *suspended load*, when particles are transported and maintained in the main body of the flow by turbulent mixing processes (Figure 3.8A). This distinction is somewhat arbitrary because there is an interchange of particles between the two modes of transport. Details of the exchange process are not well understood. The bed-material load is the principal concern because of its influence on the adjustment and development of river channel form.

The dissolved load

Solutes in rivers are derived from rock and soil weathering, from the atmosphere, and from the effects of man's activity, the last of which is becoming increasingly important as the discharge of industrial effluent and the use of agricultural fertilizers increase. Much of the dissolved load is supplied by some form of subsurface flow because contact times between water and soluble materials are longer. Consequently basins where subsurface flow is a major contributor to stream discharge and where materials are readily soluble tend to have higher dissolved loads.

Unlike suspended sediment, solute concentration (C) declines with increasing discharge (Q) at a cross-section to give an n value in

$$C = kQ^n \qquad (3.14)$$

which is usually less than 0. This reflects a dilution effect as the contribution from near-surface flows with lower solute levels increases at higher discharges. However, since n usually exceeds -1, the total dissolved load ($Q_{diss} = C.Q$) continues to increase with discharge. Indeed the dilution effect may become progressively less

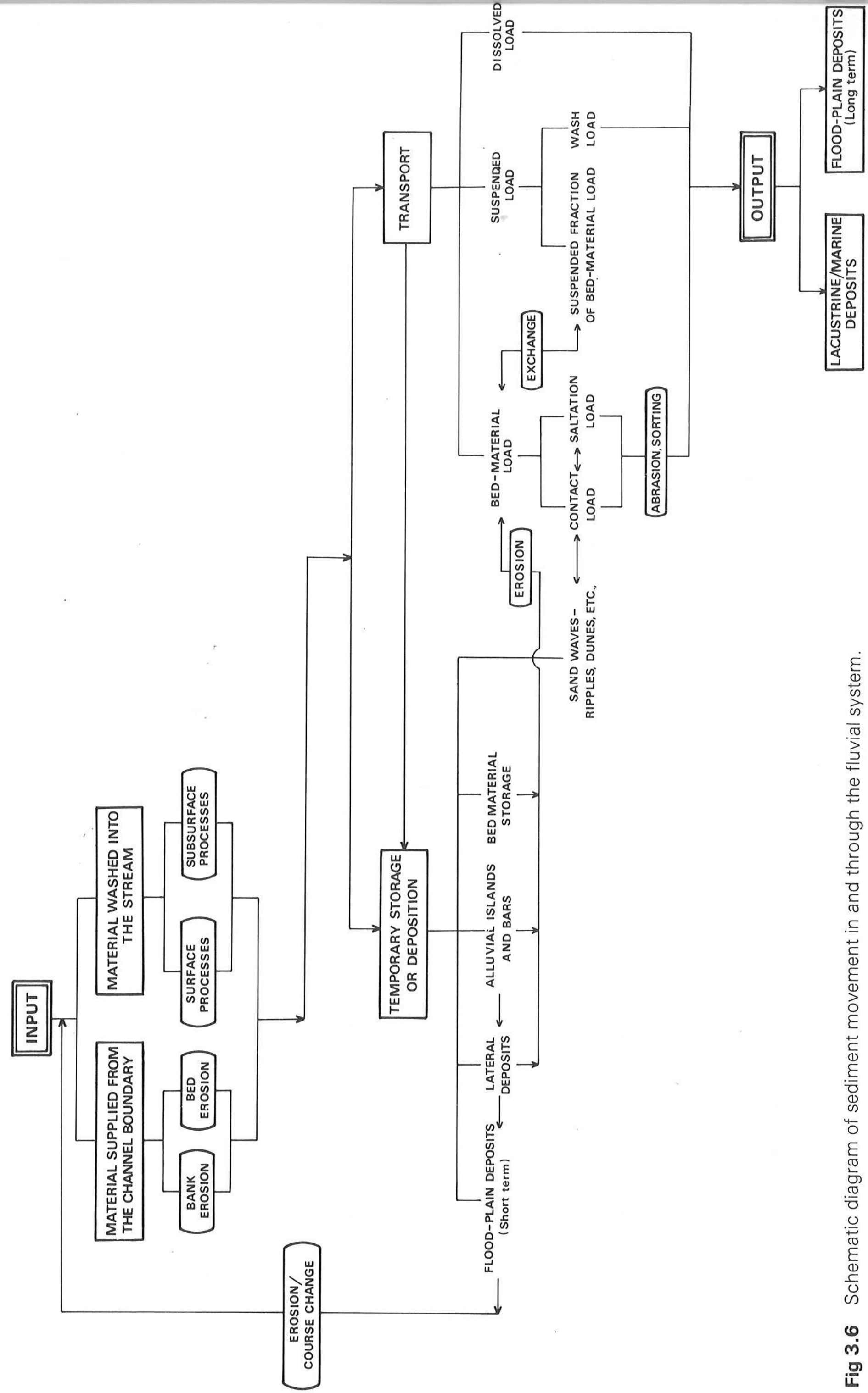


Fig 3.6 Schematic diagram of sediment movement in and through the fluvial system.

Table 3.6 Sediment and dissolved loads of major rivers

River	Mean discharge (10^3 m^3 s^{-1})	Drainage area (10^3 km^2)	Mean sediment load (10^6 tonnes yr^{-1})	Mean dissolved load (10^6 tonnes yr^{-1})	Per cent of total load carried in solution
Africa					
Congo	39.2	4 000	53	47	48
Zambezi	7.1	1 340	100	15	13
Niger	6.1	1 125	68	10	13
Orange	2.9	1 000	150	12	7
Nile	2.8	3 000	111	17	14
Asia					
Brahmaputra	19.3	580	795	75	12
Mekong	18.3	795	346	59	14
Yenisei	17.2	2 600	13	73	83
Lena	16.3	2 430	15	85	83
Ganges	11.6	975	524	76	13
Huang Ho	1.5	752	1 600	–	–
Australia					
Murray-Darling	0.7	1 070	32	9	22
Europe					
Volga	8.4	1 350	26	77	77
Danube	6.4	805	68	60	48
Dnieper	1.6	500	1	11	91
North America					
Mississippi	18.4	3 267	350	131	27
St Lawrence	10.7	1 025	5	54	91
Mackenzie	9.6	1 800	117	70	37
Columbia	8.0	670	29	35	56
Yukon	6.2	770	79	34	30
South America					
Amazon	175	6 300	498	290	37
Orinoco	30	950	86	50	37
Parana	18	2 800	112	56	33
Magdalena	7.5	240	240	28	10

Sources: Holeman (1968), Inman and Nordstrom (1971), Meybeck (1976)

significant at very high discharges when the concentration is dominated by the near-constant solute content of the storm runoff component (Gregory and Walling, 1973). The variation of solute concentration both within and between storms tends to produce a wide scatter of points on concentration–discharge graphs.

A large part of the dissolved load is carried by relatively frequent flows. In the Mississippi about 90 per cent is transported by flow events occurring monthly (Sedimentation Seminar, 1977). Dissolved load makes a significant contribution to the total load, implying that solution is an effective denudational process, although measured values may include the effects of man's activities. Accurate data are rather sparse but it has been estimated that about 41 tonnes km^{-2} of dissolved material are carried each year to the oceans (Meybeck, 1976), comprising roughly 38 per cent of the total load of the world's major rivers. However, the proportion of the total load carried in solution and the actual amounts vary considerably (Table 3.6). Also, non-denudational sources such as rainfall inputs and pollution can significantly affect the figures. Those factors which influence the relative contributions of surface and subsurface flow to total stream discharge largely control the sediment–solute balance. Dissolved load is less dependent than sediment load on the quantity of flow and, since it has little effect on stream behaviour relative to channel form adjustment, it is largely ignored in the subsequent discussion.

The wash load

The wash load moves in suspension at approximately the same speed as the flow and only settles out where flow velocities are much reduced. Some very large volumes are involved (Table 3.6), particularly along the Huang Ho where most of the silty load is derived from erosion of the loess lands in central China (Stoddart, 1978).

The rate of wash load transport is principally determined by its rate of supply from the drainage basin rather than the transport capacity of the stream. Most of the material is supplied from the erosion of cohesive river banks, with fine sediment being sheared off by the flow or thrown into suspension after bank collapse, and from surface and subsurface erosion in the catchment area by such processes as rainsplash and surface wash. While bank erosion is partly dependent on flow characteristics, the second source is independent of conditions in the stream.

The relative contributions from channel and non-channel sources vary with basin size. Although difficult, their assessment can be approached by: direct measurement of source area erosion, which is feasible only in small basins; comparison of measured sediment loads with yields estimated from a soil loss equation and a delivery term; or measurement of selected suspended sediment properties. Grimshaw and Lewin (1980) distinguished channel from non-channel sources on the basis of sediment colour in the River Ystwyth of central Wales. In this small basin (A_d = 170 km^2), just over half of the suspended sediment came from the channel itself but the proportion of non-channel sediment increased at very high discharges. In general, the upper parts of catchments which have steeper and shorter slopes tend to supply sediment from non-channel sources. Further downstream where slopes are longer and less steep, the potential for temporary storage of eroded material increases and the contribution from channel erosion becomes relatively more important. At least 65 per cent of the material carried by the lower Waimakariri River in New Zealand is supplied locally from the bed and banks of the channel (Griffiths, 1979).

Being supply- rather than capacity-limited, the rate of wash load transport is not directly a function of stream discharge. Because of the highly variable character of sediment supply, plots of wash load concentration against discharge often show a very wide scatter of points and no well-defined relation (e.g. Colby, 1963). Part of that scatter may be the result of hysteresis in which larger loads occur on the rising rather than the falling stage at the same discharge. Sediment yields from the catchment are usually highest soon after the start of rainfall when sediment is more readily available for transport, so that most of the sediment supplied by surface wash reaches the stream when discharge is rising. However, hysteresis relationships are not independent of basin size. In small basins the sediment peak tends to precede the discharge peak but may lag behind the discharge peak by a considerable time in large basins where upstream sources continue to supply the bulk of the load.

Despite the attendant problems, sediment-rating curves relating suspended sediment load (Q_{susp}) to discharge (Q) in the form

$$Q_{susp} = rQ^j \tag{3.15}$$

are often used in analysing sediment transport characteristics (Figure 3.9A). Values of j typically lie in the range of 1.5 to 3. However, a clear distinction is not always drawn between the wash load and the suspended fraction of the bed-material load in constructing such curves. The first is largely controlled by the availability of sediment in the catchment, while the latter is strongly related to flow conditions. Consequently estimates of catchment erosion from such equations are liable to error unless account is taken of different supply sources and basin size.

Sediment yield, or the total sediment outflow from a watershed, includes wash load as the dominant component. It is controlled by four main groups of factors: precipitation and runoff characteristics; soil resistance; basin topography; and the nature of the plant cover. Its overall variation with climate is commonly defined by curves which relate sediment yield to mean annual precipitation or mean annual runoff (Figure 3.7). The Langbein and Schumm (1958) curve based on group-averaged data reaches a peak at an effective precipitation of about 300 mm, trailing off at lower values because of lower runoff totals and at higher values because an increasingly abundant vegetation cover affords better protection against erosion.

The broad pattern indicated by that curve needs qualifying in several respects. Dendy and Bolton (1976), again using United States data, produced a curve in which the sediment yield peak occurs at a slightly higher precipitation of 450–500 mm. However, tropical areas are not represented in either data set and it seems that sediment yield may again begin to increase for precipitation amounts over 1100 mm, reaching a second maximum at 1200–1500 mm in humid climates with highly seasonal rainfall (Figure 3.7B and C). Indeed sediment yield may be more a function of the seasonal variability of precipitation than of the annual amount (Wilson, 1973). A comprehensive reanalysis of available data by Walling and Kleo (1979) has revealed not only considerable variability in annual sediment yields but also a more complex average relationship with three peaks (Figure 3.7D). The first could reflect the simple relationship proposed by Langbein and Schumm (1958), while the remaining two may indicate the effects of seasonal precipitation regimes associated with areas of high rainfall mediterranean climate (1250–1350 mm) and tropical monsoon conditions (> 2500 mm). In effect the Langbein and Schumm curve incorporates the interacting factors of precipitation/runoff magnitude and vegetation cover, but the former may be more significant than those authors indicated since the increasing vegetation cover in more humid areas

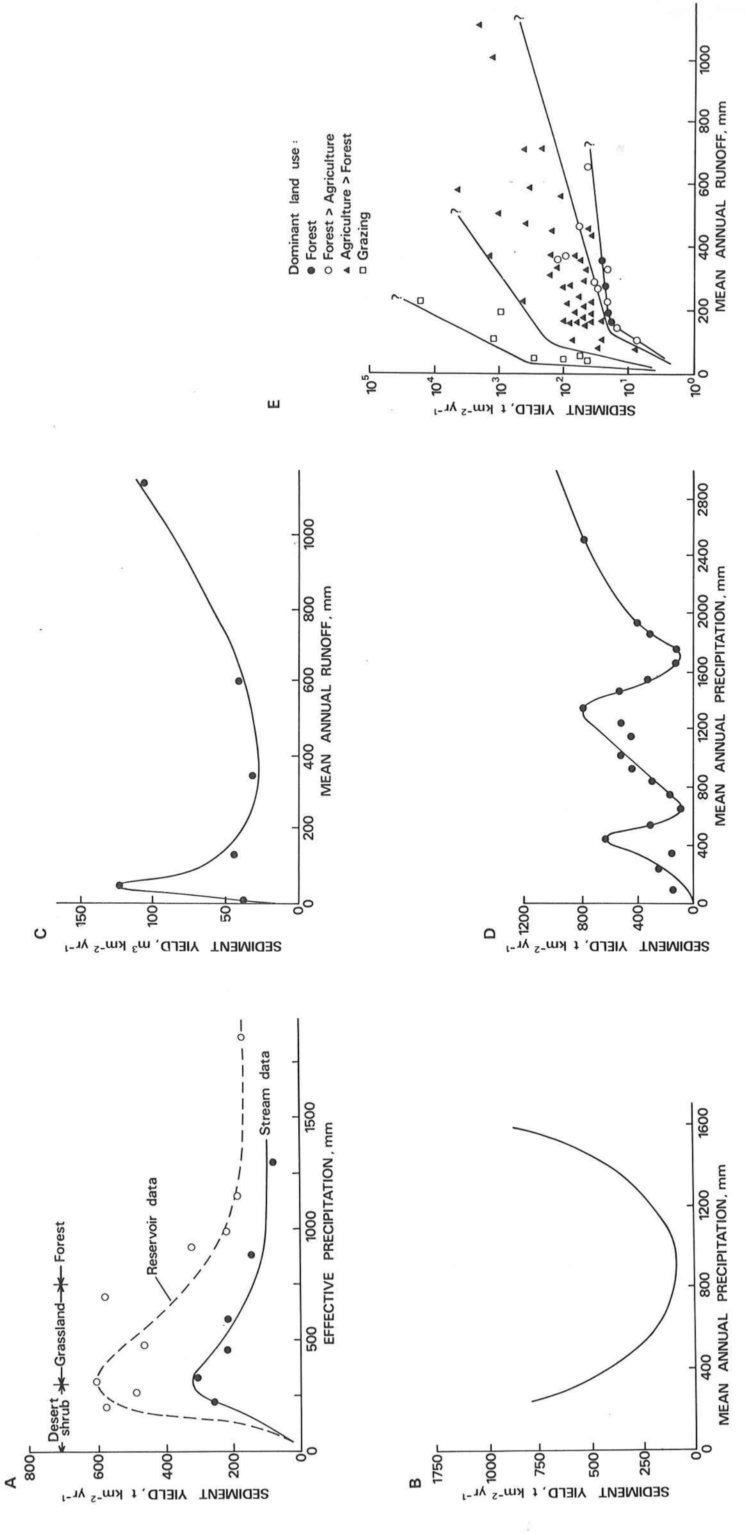


Fig 3.7 Relationship of sediment yield to:
A. Effective precipitation (after Langbein and Schumm, 1958);
B. Mean annual precipitation (after Fournier, 1960);
C. Mean annual runoff (after Douglas, 1967);
D. Mean annual precipitation (after Walling and Kleo, 1979);
E. Mean annual runoff for four land use types (after Dunne, 1979).

does not reduce erosion to the extent originally supposed. Also, it should be recognized that the influence of relief (which may be predominant according to some authors) and the widespread impact of human activity serve to distort any global pattern and especially one which is based on precipitation magnitude alone.

Although implicitly included in Langbein and Schumm's curve, the effect of vegetation or ground cover cannot be expressed by any single relationship between sediment yield and a simple climatic index in view of land use variability even within relatively uniform climatic areas. A 9-year study in small Colorado catchments showed that the conversion from sagebrush to grass cover reduced sediment yield by 80 per cent without significantly affecting annual runoff totals (Lusby, 1979). Dunne (1979) has analysed the sediment yields of 61 Kenyan catchments classified according to the land-use types:

(i) Forest (F),
(ii) Forest cover > 50 per cent of the basin with the remainder under cultivation (FA),
(iii) Agricultural land > 50 per cent with the remainder forested (AF),
(iv) Grazing land (G).

Within each type sediment yield increases with mean annual runoff (related to mean annual precipitation by $Q = 0.000033P_m^{2.27}$), but major differences exist in the rate of that increase (Figure 3.7E). As the average cover density decreases from (i) to (iv), sediment loss becomes increasingly sensitive to runoff and there appears to be a progressive increase in the rate at which sediment yield varies. Dunne concluded that sediment yield, which varied considerably from 8–20,000 t km^{-2} yr^{-1}, was largely controlled by land use, with climatic and topographic factors having subsidiary effects. These results emphasize the potential variability of sediment yield even within relatively small areas and the need for caution when using global syntheses, particularly in palaeohydrologic reconstructions (chapter 5).

Although suspended sediment transport is more erratic than dissolved load transport and is mainly associated with higher discharges, most of the material is still carried by relatively frequent events of moderate magnitude. For various rivers Wolman and Miller (1960) calculated that about 99 per cent of the total suspended load is transported by flows recurring more frequently than once every 10 years, and that 80–90 per cent of the load is carried by flows recurring more frequently than once a year. Despite subsequent modifications (e.g. Baker, 1977), evidence points to the validity of these conclusions for a wide range of conditions (Sedimentation Seminar, 1977; Webb and Walling, 1982). Catastrophic events may individually carry larger amounts of sediment but their contribution to suspended sediment transport over the entire range of discharges is less significant because they recur so infrequently.

Catchment erosion represents not only a loss of assets to agriculture and forestry but also a potential liability further downstream. Reservoir siltation reduces storage capacity and the life expectancy of reservoirs, while the silting-up of rivers may affect navigation and increase the risk of flooding in lower reaches. The silt concentration in the Huang Ho is so high that a reduction of 10–40 per cent could decrease peak discharge by about 20 per cent without any change in the actual volume of water (Stoddart, 1978).

Wash load transport has effects of more immediate relevance to stream behaviour and channel form adjustment. Very high concentrations damp down turbulence and alter the apparent viscosity of the flow, enabling the transport of a slightly larger bed-material load than would otherwise be the case (Simons *et al.*, 1963). Where there is strong infiltration from the stream into the bed and banks, as in ephemeral streams, part of the wash load may be deposited as a caked layer which stabilizes the boundary and improves its resistance to erosion (Harrison and Clayton, 1970). Large suspended loads can also contribute to rapid changes in stream course, and influence the style of flood-plain development.

On the whole, the dissolved and wash load components have little direct influence on channel geometry. However, rivers with a large wash load may be morphologically different from those with a relatively large bed-material load because of the long-term effect of transport characteristics on the boundary composition of alluvial channels. A large wash load is in part diagnostic of a high silt–clay content in the channel banks. Using the percentage silt–clay in the channel perimeter as an index (M) of the type of sediment load, Schumm (1960) has argued that streams carrying predominantly wash load should have channels which are relatively narrow and deep. A larger bed-material load requires wider and shallower cross-sections in which greater shear stresses are directed against the bed, enabling the stream to transport the coarser load.

The bed-material load

Unlike dissolved and wash load transport, the rate of bed-material transport is almost entirely a function of the transporting capacity of the flow. Field measurements are notoriously difficult to make and much work has been carried out in laboratory flumes where conditions can be simplified. Many variables are involved (Table 3.7), whose independence/dependence cannot always be determined as simple cause and effect because of intercorrelation. Indeed, the independence or dependence of variables alters with the system of interest, be it flumes, natural streams in the short term or natural streams in the long term. The basic problems are twofold: to understand the dynamics of bed-material movement; and to establish a relationship between sediment transport rate and relevant properties of the flow, fluid and sediment, thereby enabling the prediction of transporting capacity for given conditions.

Table 3.7 Variables pertinent to bed-material transport

Flow properties	Fluid properties	Sediment properties	Other properties
Discharge (Q)	Kinematic viscosity (ν)	Density (ρ_s)	Gravity (g)
Velocity (v)	Density (ρ)	Size (D)	Plan-form geometry
Flow depth (d)	Temperature (T)	Sorting (σ)	
Width (w)	Wash load concentration (C)	Fall velocity (v_s)	
Slope (s)			
Resistance (ff)			

As regards the **dynamics of movement**, particles roll, slide or saltate along the bed in a shallow zone only a few grain diameters thick once sediment motion begins ($\tau_o > \tau_{cr}$). Material transported in this way constitutes the *bed load*. Rolling is the primary mode of transport in gravel-bed streams, while saltation is largely restricted to grains of sand size (Table 3.3). With further increases in the strength of the flow, the less massive particles may be carried upwards into the main body of the flow to be transported as *suspended load*, possibly once a second threshold ($\tau_o > \tau'_{cr}$ where $\tau'_{cr} > \tau_{cr}$) has been reached (Yalin, 1972). Movement in suspension is maintained against gravity by turbulent eddies which are random in strength and

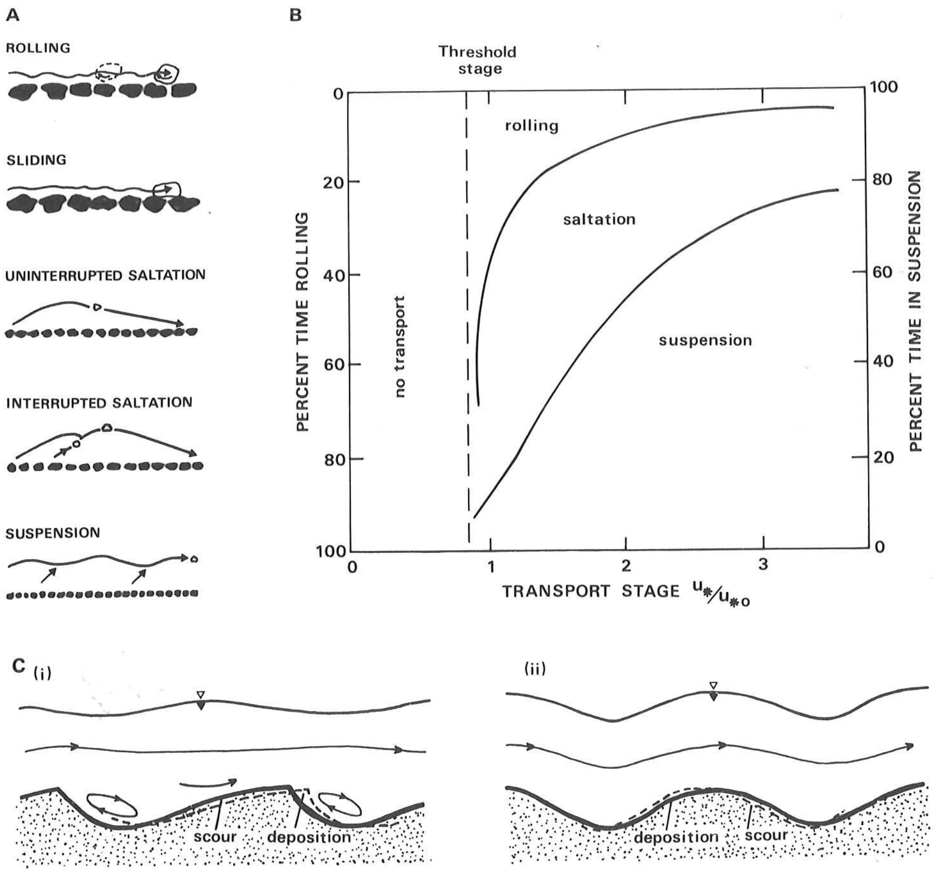


Fig 3.8 A. Modes of transport of bed-material load.
B. Percent of time single particles in a water stream experience rolling, saltation and suspension as a function of transport stage. Percent of time in saltation is represented by the distance between the curves (after Abbott and Francis, 1977).
C. Schematic diagram of the pattern of erosion and deposition over a dune (i) and antidune (ii) bed.

direction, so that grains do not follow predictable paths. The conventional separation of bed-material load into these two components is an idealization. A distinct value of τ'_{cr} corresponding to the initiation of suspended transport does not exist in nature; nor is there a sharp boundary separating regions of bed load from suspended load transport.

The theoretical formulation of a transport function requires a set of criteria with which to distinguish the several modes of transport (rolling, saltation and suspension; Figure 3.8A). Based on flume experiments in which the movement of single grains was traced photographically, Abbott and Francis (1977) attempted to define such a criterion in terms of transport stage, u_*/u_{*o}, where u_* ($= \sqrt{\tau_o/\rho}$) is the shear velocity of an observed flow and u_{*o} is the threshold shear velocity at which motion begins. The development of suspension from saltation (regarded as an intermediate step) occurred much less rapidly than that of saltation from rolling (Figure 3.8B). Rolling soon gave way to saltation, the time spent in the rolling mode

decreasing very rapidly from about 60 per cent close to the threshold stage ($u_*/u_{*o} = 1$) to 20 per cent at $u_*/u_{*o} \sim 1.4$. Saltation was a persistent mode of transport even above $u_*/u_{*o} = 1.9$ when suspension became dominant. Figure 3.8B suggests that all modes of bed-material transport can occur simultaneously over a wide range of flow conditions, even though one may predominate at any given stage. Although higher stresses are probably required for full suspension in natural rivers, this work demonstrates how the transport modes in sandy beds can be distinguished. It does, however, exclude the effects of grain – grain interactions which become important when $u_*/u_{*o} > 2$ (Leeder, 1979).

Grain movement is characteristically intermittent, especially in gravel-bed streams when the flow is just above the erosion threshold. In sand-bed streams where ripples and dunes develop, grains tend to move in groups as the bed form migrates downstream (Figure 3.8C). Grains on the lee side may be temporarily buried and not re-exposed until the bed form has progressed. The development of such forms emphasizes the strong links between sediment transport, bed configuration, resistance and flow conditions close to the bed. The irregularity of movement in which individual particles travel short distances before temporarily coming to rest has led to the formulation of stochastic models, with appropriate probability distributions being defined for the step lengths and intervening rest periods.

Sediment transport equations give the maximum amount of material (capacity) that can be carried for given conditions of the flow, fluid and sediment. In parallel with the subdivision of bed-material load, separate equations have been developed for bed load, suspended load and total load using both deterministic and probabilistic approaches.

Many *bed load formulae* can be classified according to whether they relate the sediment transport rate per unit width (q_{sb}) to either excess shear stress ($\tau_o - \tau_{cr}$) or excess discharge per unit width ($q - q_{cr}$):

$$q_{sb} = X' \tau_o (\tau_o - \tau_{cr}) \qquad \text{Du Boys type} \qquad (3.16)$$

$$q_{sb} = X'' s^k (q - q_{cr}) \qquad \text{Schoklitsch type} \qquad (3.17)$$

where X′ and X″ are sediment coefficients (Graf, 1971). Based on his view of the stream as a transporting machine which expends power to perform work, Bagnold (1977) has developed a semi-theoretic bed load function,

$$q_{sb} \simeq (\omega - \omega_o)[(\omega - \omega_o)/\omega_o]^{\frac{1}{2}}(d/D)^{-\frac{2}{3}} \qquad (3.18a)$$

which is essentially of the second type (3.17) with an added relative roughness term (d/D) and stream power per unit width ($\omega = \gamma Qs/w$) replacing discharge. Although the function has subsequently been modified in two main respects to give (Bagnold, 1980)

$$q_{sb} \simeq (\omega - \omega_o)^{\frac{3}{2}} d^{-\frac{2}{3}} D^{-\frac{1}{2}} \qquad (3.18b)$$

it still implies a greater transport rate not only at a higher excess power $\omega - \omega_o$ (constant depth and grain size) but also at a lower flow depth d (constant excess power and grain size) (Figure 3.9B). The problem remains, however, of determining the threshold stream power ω_o which is not directly measurable in natural rivers.

Einstein's (1950) approach represented a break with traditional methods in that explicit recognition was given to the stochastic nature of bed particle movement

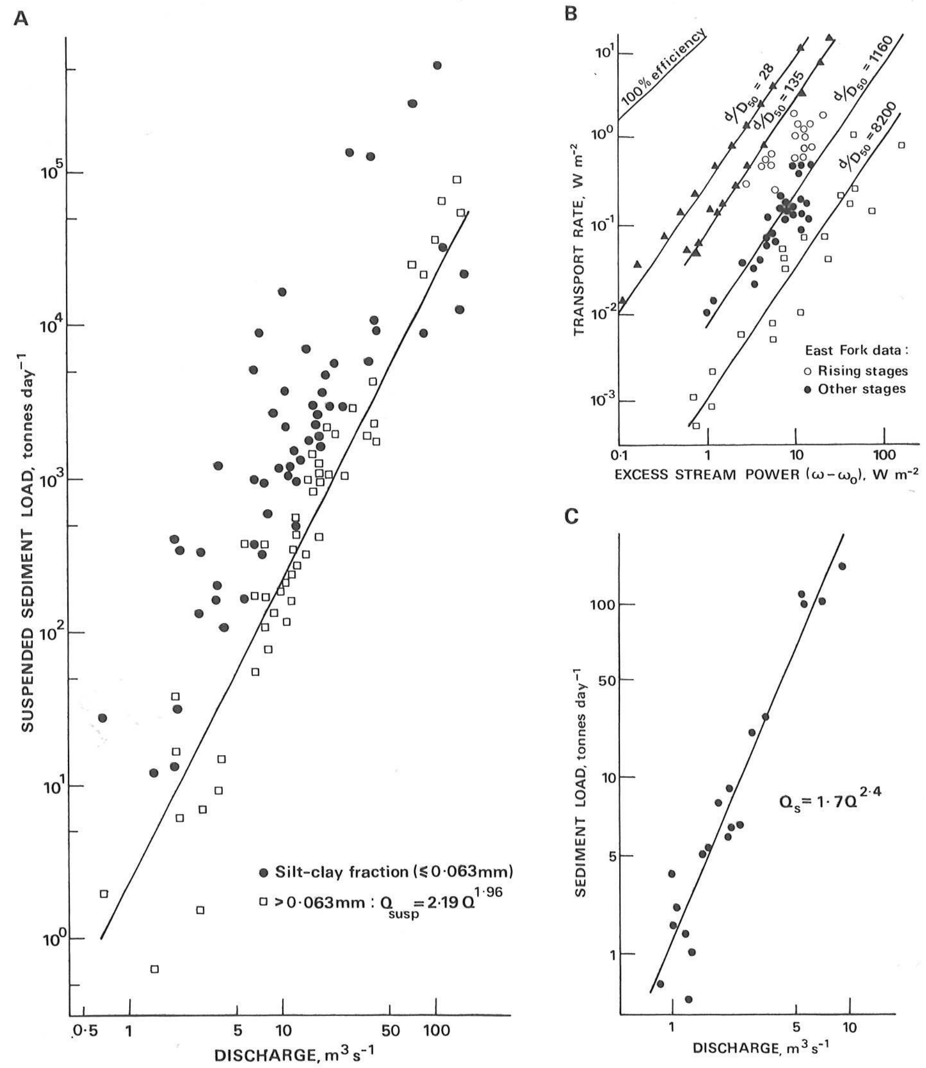


Fig 3.9 A. Variation of suspended sediment load with discharge for grain sizes less than and greater than 0.063 mm, Rio Grande at Albuquerque (data of Nordin and Beverage, 1965). The plotted line applies only to the coarser fraction.
B. Bed-load transport rate as a function of excess stream power per unit bed area ($\omega - \omega_o$) for different relative depths (d/D_{50}) (after Bagnold, 1977). The data include measurements made in a laboratory flume (▲) and natural rivers (○, ●: East Fork River; □: Clearwater River).
C. Sediment load as a simple power function of discharge, River Bollin.

and to the different rates at which individual size fractions in the load may be transported. The transport rate

$$q_{sb} = NW \tag{3.19}$$

is the product of the weight of each particle (W) and the number of particles (N) eroded from an area of the bed (A), where

$$N = \frac{Ap}{a} \tag{3.20}$$

p is the probability of any particle being eroded per unit time and a the bed area occupied by each particle. Clearly the problem lies in determining p.

Many bed load equations rely on the empirical determination of coefficients, which limits their overall applicability. The equations predict the maximum load that a stream can theoretically carry but capacity transport requires an unrestricted supply of material. In heterogeneous beds the flow may be closer to capacity for one size of material than for another, since transport efficiency declines with increasing grain size (Leopold and Emmett, 1976). If only part of the material is moved over a wide range of flows, the coarser fraction may accumulate at the surface to form a protective armour coat which hinders or prevents further transport. Such time-based elements of the transport process are geomorphologically relevant but are not incorporated in conventional equations.

The *suspended load* (q_{ss}) can be calculated from

$$q_{ss} = \int_o^y cv \quad dy \tag{3.21}$$

provided the vertical variation of velocity (v) and suspended sediment concentration (c) with depth (y) is known. Most analytical treatments of suspension are based on the concept of diffusion whereby particles are dispersed in the flow through the action of turbulent mixing. In general, diffusion models describe a vertical distribution of suspended sediment in which both the concentration and grain size decrease exponentially with distance from the bed. Although the suspension process is not completely understood, agreement between predicted and observed values seems to be much better for suspended than for bed load transport.

At low transport rates most material moves close to the bed and the *total bed-material load* can be estimated from bed load equations. At other times the total load ($Q_s = q_s.w$) can be obtained in one of two ways: indirectly, by simple addition of the two components estimated separately,

$$Q_s = Q_{sb} + Q_{ss} \tag{3.22}$$

or directly, from a total load formula in which the need to distinguish between the bed (Q_{sb}) and suspended (Q_{ss}) loads is avoided. Some procedures include wash load while others do not, depending on whether or not they rely on field measurements of suspended sediment.

The accuracy of transport equations is difficult to check because reliable measurements of bed-material discharge are relatively scarce. Considerable errors can be introduced if corrections are not made for the wash load component in suspended sediment samples which are often used as the basis for estimates (Graf, 1971). Despite the problem of evaluating equations developed initially for different ranges of flow and sediment conditions, extensive tests have been carried out (Task Committee, 1971a; White *et al.*, 1973). The large variation in results between different formulae and the wide divergence between predicted and measured loads point to the general inadequacy of sediment transport theory. There is no general agreement on a suitable list of independent variables for predicting sediment discharge. Indeed such is the range of factors involved (Table 3.7) that a widely acceptable theory will be slowly if ever achieved (Vanoni, 1975). Simple relationships of the form,

$$Q_s = kQ^n \tag{3.23}$$

may suffice as first-order approximations to observed data (Figure 3.9C; Task Committee, 1971a).

The bed-material transport rate can be highly variable both within and between cross-sections in a reach, and with time. Fluctuations may be random or quasi-periodic, and of different time scales. Over intervals measured in minutes, fluctuations occur as dunes pass a given point, maximum amounts of transport being associated with the passage of dune peaks and smaller amounts with that of intervening troughs (Leopold and Emmett, 1976). In gravel-bed streams, large bars may take a full season to move through a cross-section. These fluctuations emphasize the need for sampling over long enough time periods in order to obtain reliable time-averaged estimates of sediment discharge.

The transport rate may vary during the hours, days or months that a flood wave takes to move through a channel reach, leading to short-term changes in stream bed elevation as a result of *scour and fill*. It has been observed that in sand-bed streams particularly the channel bed is scoured at high discharges on the rising stage and filled to approximately the pre-flood level on the falling stage (Leopold *et al.*, 1964), and that scour occurs more or less continuously along a reach (Emmett and Leopold, 1965). However, the spatial and temporal pattern of the scour-and-fill process is more variable than these statements would suggest. Specific loci of scour exist, in channel bends, at tributary junctions and around obstructions. Scour and fill may alternate locally several times during a flood, with only a small part of a reach experiencing simultaneous scour or fill at any one time, implying that the process may be related to bed-form migration (Foley, 1978). Andrews's (1979) observations along a short reach (430 m) of the East Fork River revealed two distinct types of cross-section:

(i) 'scouring' sections which scoured at discharges (Q) above bankfull (Q_b) and filled at $Q < Q_b$, when the sediment-transport rates were respectively relatively large and small; and
(ii) 'filling' sections which filled at $Q > Q_b$ and scoured at $Q < Q_b$, when the respective transport rates were relatively small and large.

At any discharge except bankfull, some sections were accumulating bed material (fill), while others were losing bed material (scour). The sequence of scour and fill was related to the distinctive hydraulic geometries of the two types of section (Table 3.8), in which significant reversals occurred at about bankfull in the relative magnitude of variables (notably velocity) because of their different rates of change in scouring and filling sections. Interestingly the channel had returned to its pre-flood elevation within one year. Thus the scour-and-fill process may be regarded as a short-term mechanism for smoothing out irregularities in the transport rate and thereby maintaining a channel reach in quasi-equilibrium.

In the short span of engineering time, the sediment transport rate can be

Table 3.8 Hydraulic characteristics of scour and fill sections, East Fork River (after Andrews, 1979)

Characteristic	Scour sections	Fill sections
b	$< \bar{b}$	$> \bar{b}$
f	$\sim \bar{f}$	$\sim \bar{f}$
m	$> \bar{m}$	$< \bar{m}$
p	$< \bar{p}$	$> \bar{p}$
j	$> \bar{j}$	$< \bar{j}$
Channel shape	Relatively narrow and deep	Relatively wide and shallow
Velocity	Larger velocity at higher discharges	Larger velocity at lower discharges

Symbols: b, f, m, p and j are respectively the rates of change of width, depth, velocity, resistance (ff) and bed-material transport with discharge; $\bar{b}$, $\bar{f}$, $\bar{m}$, $\bar{p}$ and $\bar{j}$ are the corresponding reach averages obtained from 11 cross-sections.

regarded as a variable dependent on flow, fluid and sediment properties (Figure 1.1), provided any differences in the input and output of sediment along a reach are temporarily accommodated by scour or fill. In the longer term a stream adjusts its geometry to the water and sediment discharges supplied. Equation (3.18) implies that, if over a period of time a stream is neither aggrading nor degrading its channel, the width/depth ratio must have become adjusted to maintain a continuity of sediment transport in which input and output are equal. The bed is then protected from downcutting by a constant sediment layer. Any increase or decrease in the mean supply rate of mobile sediment would result in aggradation or degradation in an attempt to restore a balance by respectively decreasing or increasing relative roughness d/D, which would have the reverse effect on the mean transport rate (Bagnold, 1977).

Most rivers have had hundreds of years in which to make this adjustment. However, sediment continuity can be upset by man's activities. Reservoirs are very efficient sediment traps and a stream may degrade its bed below a dam in order to adjust to the larger discharge/load ratio. Channel improvements such as straightening and bank strengthening may so increase flood heights that downstream transport is severely disrupted. The important point is that a close relationship exists between channel form and the input–output conditions of sediment load in a channel reach. The absence of a general bed-material transport theory is a drawback in modelling that relationship adequately.

One significant aspect of the relationship is the relative effectiveness of events having different magnitude–frequency characteristics. Although based initially on suspended load measurements, the view that relatively frequent events of moderate magnitude perform most of the work (Wolman and Miller, 1960) is supported by evidence from streams which carry predominantly bed load (Pickup and Warner, 1976; Andrews, 1980). In the Yampa River of Wyoming, Colorado, the most effective discharges in terms of bed-material transport recur on average between 1.5 and 11 days each year (Figure 4.3D; Andrews, 1980). However, only about 25 per cent of the bed material is then in motion, so that coarser fractions tend to accumulate and form a stabilizing layer. Catastrophic events, particularly in small basins, may redistribute larger material to such an extent that channel form is more or less permanently affected because that material remains immobile under subsequent flow conditions (Stewart and La Marche, 1967).

Downstream changes in bed material characteristics represent an element of the transport process which is of immediate geomorphological relevance. Since several of the factors, notably slope and discharge, which influence the rate of particle movement change systematically in the downstream direction, both the size and size distribution (sorting) of bed material may be expected to vary similarly. Two sets of processes are involved: abrasion and sorting.

Abrasion is a summary term covering such mechanical processes as grinding, impact and rubbing, which chip and fracture particles not only during transport but also in place when the combined effects of lift and drag forces cause particles to vibrate (Schumm and Stevens, 1973). Sorting reflects the action of:

(i) selective entrainment, in which only that fraction of the bed material smaller than the threshold size is transported by a given flow event (Figure 3.5C and D); and

(ii) differential transport, in which smaller particles are transported faster and further than are larger ones.

Clearly sorting is influenced by all the variables involved in sediment transport. Chemical breakdown may also contribute to downstream changes under certain conditions.

Historically, abrasion was regarded as the dominant process responsible for the size reduction of bed material commonly observed along alluvial rivers. However, largely as a result of abrasion tank experiments (e.g. Kuenen, 1956), it was realized that observed rates of reduction were much larger than could be attributed to abrasion effects. Bradley *et al.* (1972) concluded that about 90 per cent of the downstream size reduction in Knik River gravels is caused by sorting, with the balance attributable to abrasion. However, recent work has re-established the status of abrasion (in conjunction with in situ weathering) as an effective process of particle wear (Schumm and Stevens, 1973), with maximum effectiveness possibly in upper reaches where slopes are steeper and particles larger.

Several models have been developed to describe the two processes. If, as a result of abrasion, it is assumed that the decrease in particle size (D) per unit distance downstream (L) is proportional to particle size,

$$\frac{dD}{dL} = -\alpha D \qquad (3.24)$$

then the solution describes an exponential decrease of particle size with distance,

$$D = D_o e^{-\alpha L} \qquad (3.25)$$

where α is a coefficient of abrasion and D_o is the initial grain size at $L = 0$. Models of the sorting process based on standard sediment transport equations (Rana *et al.*, 1973; Deigaard and Fredsøe, 1978) or stochastic particle motion (Troutman, 1980) yield a similar result. In essence α can be thought of as including the undifferentiated effects of both abrasion and sorting. The rate of change (α) may vary over large river distances to give at least two segments, where the upper and lower segments have respectively larger and smaller rates of particle size decrease (Rana *et*

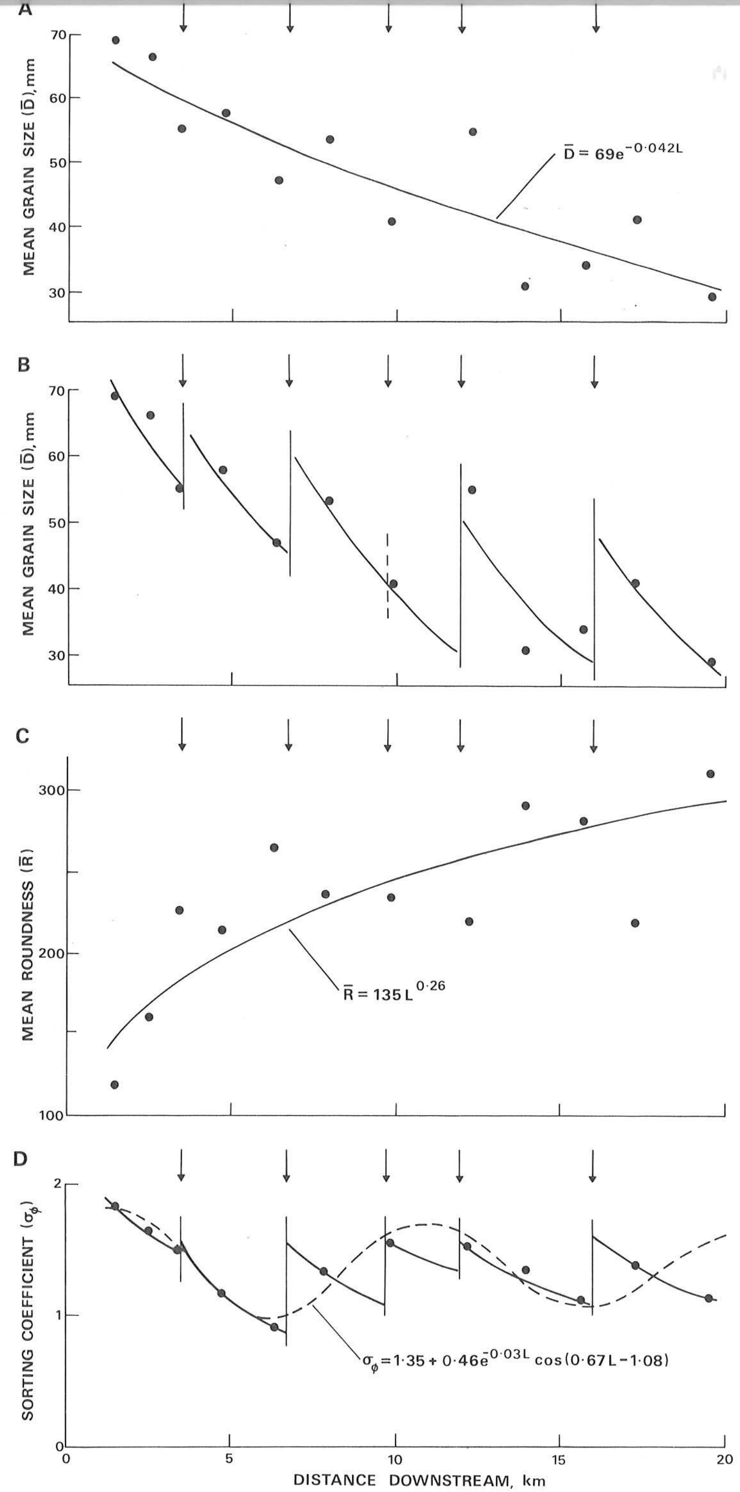


Fig 3.10 Downstream changes in bed material characteristics, River Noe, Derbyshire:
A. Exponential decrease of mean grain size;
B. Discontinuous exponential decrease of mean grain size to show the effect of tributary inflow;
C. Increase in mean particle roundness;
D. The effect of tributary inflow on sediment sorting represented as discontinuous exponential change (solid lines) or pseudo-periodic variation (dashed line).

al., 1973). Such a discontinuity has been observed in Japanese rivers corresponding to a deficiency of grains in the size range of 1–4 mm (Yatsu, 1955), which may reflect a threshold between gravel-bed and sand-bed sections of a river.

Equation (3.25) is usually calculated for the mean ($\overline{D}$) or median (D_{50}) grain size obtained from samples collected along a river by discrete or bulk sampling methods. It predicts a rapid decrease in grain size in headwater reaches where the initial material is usually coarse, and a much lower rate of change further downstream, both of which are largely borne out by observation. Grain size decreased from 330 mm to 44 mm in the first 42 km of the Knik River in Alaska (Bradley *et al.*, 1972). Only a slight reduction was observed along 3200 km of the lower Amazon where cross-channel variations in particle size at several sections were at least as great as downstream ones (Nordin *et al.*, 1980). Even in relatively short streams a downstream decrease is not always observed (Hack, 1957; Brush, 1961).

Plots of grain size against distance downstream also tend to show large amounts of scatter about regression lines (Figure 3.10A), which can be attributed to several causes. Sampling errors and the natural variability of bed material size within local reaches can introduce considerable noise into the data sufficient to mask any systematic downstream trend (Church and Kellerhals, 1978). Particle lithology is not always considered but, because lithology influences the size of material supplied initially and the subsequent rate of wear, different lithologies may be expected to behave in different ways. Finally, the introduction of fresh material from bank and tributary sources complicates the overall pattern to produce, in the case of the latter, increases in grain size below junctions (Figure 3.10B). Where a sequence of tributaries enters a main stream, grain size may vary discontinuously in such a way that an exponential decrease below each junction is followed by a stepped increase at the next junction (Troutman, 1980), the magnitude of that increase being possibly related to the relative sizes of the main stream and tributary at each confluence (Knighton, 1980b). Thus, underlying the main downstream trend are random and systematic variations not explicitly catered for by equation (3.25).

Properties other than size change through downstream transport. Particles tend to become rounder as a result of abrasion (Figure 3.10C; Mills, 1979). Bed material is generally better sorted with distance downstream, although tributary inflow can again disrupt the picture (Figure 3.10D) to produce either discontinuous change similar to that proposed for particle size (Troutman, 1980) or a pseudo-periodic variation if tributary entry follows some regular pattern (Knighton, 1980b). The main point is that sediment properties are not unchanging and their observed state reflects a wide range of influences in the long-continued action of the sediment-transport system.

Particles are reduced in size by abrasion processes and assigned their position along a stream by sorting processes. Even without the complications caused by sediment inflow from different sources, the modelling of downstream transport conditions is hindered by the unknown efficiency with which sediment transport equations developed originally for cross-sectional variations can be applied three-dimensionally to the downstream case. Strict equilibrium cannot be maintained for very long where sorting processes operate alone and a downstream decrease in particle size is to be achieved, because equality of sediment input and output would require a progressive coarsening of bed material with time at a cross-section, leading to stream bed aggradation and the development of a new slope. However, although any overall reduction in particle size must be due to abrasion and breakage, the time scale for sorting processes to change bed material size may be much shorter than the time scale for bed slope adjustment (Deigaard and Fredsøe, 1978). Downstream changes are an important element of the transport process because of the relationship between sediment properties and those aspects of stream behaviour and channel form which also vary longitudinally.

Sediment deposition

The final element of the process triumvirate, deposition, has received comparatively less attention from geomorphologists and yet alluvial rivers build a wide range of depositional forms (Table 3.9). In addition to the progressive sorting which contributes to the downstream reduction in particle size, local sorting occurs over much shorter distances, related to the local distribution of stream forces. Deposition begins once the flow velocity falls below the settling velocity of a particle, which for a given particle size is less than that required for entrainment (Figure 3.5D). Settling velocity is closely related to particle size, so that the coarsest fraction in motion should be deposited first with progressively finer grains settling out as the flow velocity continues to fall. The net effect is a vertical and horizontal (downstream and transverse) gradation of sediment sizes.

The most common depositional feature is the *flood-plain* formed from a combination of within-channel and overbank deposition, although many sedimentary forms are involved (Lewin, 1978). During lateral channel migration, erosion of one bank is approximately compensated by deposition against the other, principally but not exclusively in the form of point-bars. With continuing migration, a point-bar is built streamward and also increases in height through the deposition of sediment carried onto the bar surface by inundating flows (Figure 3.11). Since the innermost parts of a point-bar are occupied by progressively less frequent flows during construction, the final stages may take considerable time, although Hickin and Nanson (1975) estimated that point-bar ridges along the Beatton River are produced gradually over a period of 27 years. Also, progressively finer sediment is carried onto the bar as it grows surfaceward to give a vertical gradation of sizes from coarsest to finest. Over a long period of lateral and downvalley shifting, the channel may occupy all positions on the valley floor, continually building the flood-plain from within-channel deposits. From measured rates of bank erosion and historical map information, Hooke (1980) estimated the time period required for a complete traverse of the 1.2 km wide flood-plain of the River Exe in Devon to be about 1300 years.

The term flood-plain implicitly includes the idea of flooding as a natural attribute of rivers. Although bankfull discharge has a variable recurrence interval its relative frequency suggests that vertical accretion could be a major process of flood-plain construction, with natural vegetation aiding the depositional process. Schumm and Lichty (1963) and Burkham (1972) have described how new flood-plains were built after destruction by major floods in as little as 20 and 50 years respectively. Vertical accretion took several forms: (i) formation of natural levées; (ii) direct deposition in the extra-channel areas; and (iii) development of channel islands which subsequently became attached to one bank (Figure 3.11), a process which has been observed elsewhere (Knighton, 1972). The sediment deposited by overbank floodwaters comes from material carried in suspension, either as wash load (silts and clays) or the finer fractions of the bed-material load (fine and medium sands). Since the transporting ability of the flow tends to decrease away