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# Geomorphic thresholds in riverine landscapes

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**SUMMARY**

1. Rivers are subject to thresholds of several types that define significant changes in processes and morphology and delimit distinctive riverine landscapes and habitats. Thresholds are set by the conditions that govern river channel process and form, amongst which the most important are the flow regime, the quantity and calibre of sediment delivered to the channel, and the topographic setting (which determines the gradient of the channel). These factors determine the sediment transport regime and the character of alluvial deposits along the channel.
2. Changes occur systematically along the drainage system as flow, gradient and sediment character change, so a characteristic sequence of morphological and habitat types – hence of riverine landscapes – can be described from uplands to distal channels. The sequence is closely associated with stream competence to move sediment and with bank stability.
3. The paper proposes a first order classification of river channel and landscape types based on these factors. The riverine landscape is affected seasonally by flow thresholds, and further seasonal thresholds in northern rivers are conditioned by the ice regime.
4. It is important to understand geomorphic thresholds in rivers not only for the way they determine morphology and habitat, but because human activity can precipitate threshold crossings which change these features significantly, through either planned or inadvertent actions. Hence, human actions frequently dictate the character of the riverine landscape.

*Keywords*: drainage basin, fluvial competence, fluvial geomorphology, river organisation, riverine thresholds

## Introduction

The morphology of alluvial river channels is the product of certain governing conditions, amongst which the most important are the quantities of water and sediments introduced into the channel, the calibre of the sediments, and the history and physiography of the landscape through which the river runs. The landscape conditions determine the sediments available to the river and they establish the overall topographic gradient down which the river flows, hence the rate of expenditure of energy by the flowing water.

These conditions were related many years ago by the American river engineer, Emory Lane (1955), who presented the qualitative relation

$$Q_s/Q \sim S/D \tag{1}$$

wherein $Q_s$ is sediment transport, $Q$ is streamflow (so the quotient is sediment concentration), $S$ is channel gradient and $D$ represents sediment calibre. Slight rearrangement yields the expression

$$Q_s \sim QS/D \tag{1a}$$

which states that sediment transport is directly related to the stream power (represented by $QS$) and inversely related to the calibre of the sediment. This statement usefully emphasises that the capacity of a stream to transport a sediment load and the competence of the stream to move particular sizes are distinct conditions. The latter may strongly influence the former, however, depending upon what material is available to the stream. The interplay of these factors produces distinctive channel morphologies by transporting and sorting sedimentary material in varying ways. The physical processes associated with

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sediment sorting create important thresholds in river channels and in the riverine landscape. Bank strength, as mediated by sediment texture and vegetation, is an additional governing condition that affects channel transitions.

Streamflow determines the scale of a river channel and, since runoff is highly correlated with drainage basin area, river channel scale varies systematically throughout a drainage basin (Leopold, 1994). Streamflow varies continuously in time according to recent weather and seasonal flow history, whereas channel morphology remains relatively fixed in the short term. Varying flows moving through a fixed channel geometry create additional thresholds of temporal significance.

In this paper, I review morphological transitions and thresholds associated with sediment and water transfers through the landscape. The major sections in the paper treat the influence of stream competence on the distribution of sediments and the character of the channel through the drainage system, and the effect of variations in sediment transport capacity on channel morphology. These considerations cover the two sediment-related governing factors in Lane's relation. The final section considers the effect of variations through time of water flow, the main driving force for the entire system. Comments will be included on the significance of these transitions for ecosystems and the effect of human activities in the riverine landscape. The 'riverine landscape' in this paper is understood to consist of the channel zone and the adjacent riparian zone, extended to the limit of influence of contemporary fluvial processes. This includes the entire active floodplain of the river.

First, drainage system functions are reviewed, for this gives rise to the most significant transitions of all.

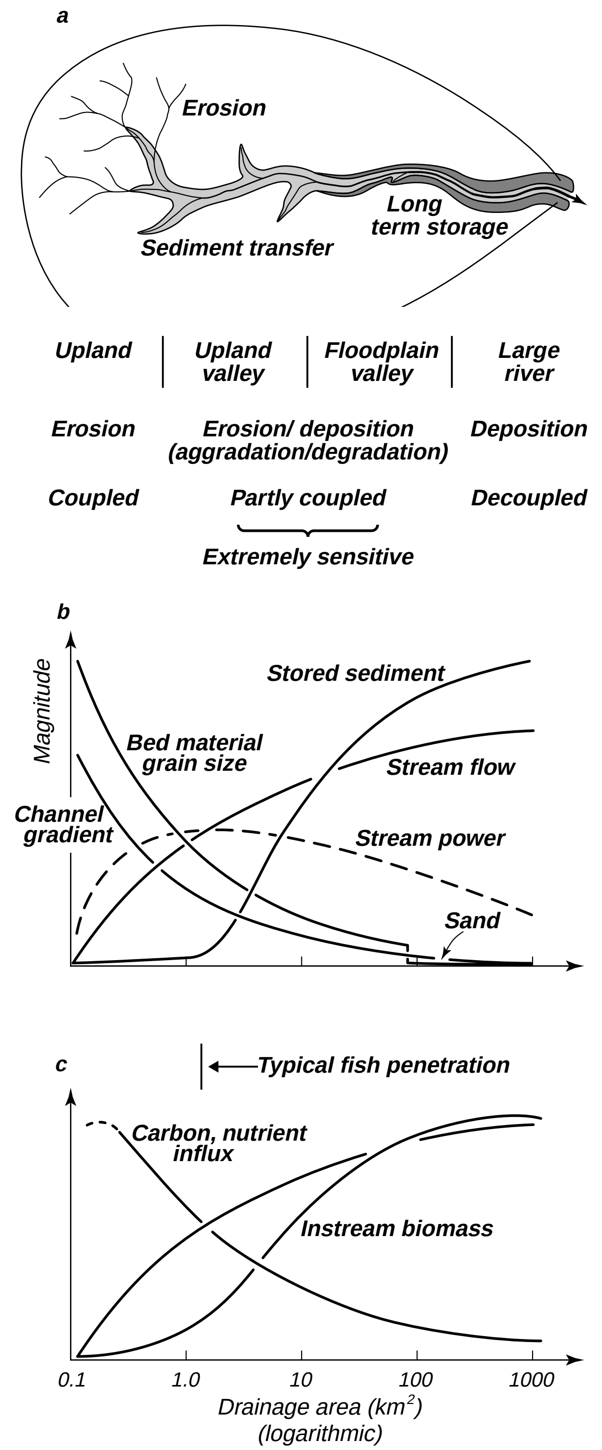
## Drainage basin structure and function

As water runs off the land, it collects into stream channels which combine in a treelike network. The confluence of tributaries downstream and the steady increase in area drained produce stepped increases in the flows and size of the channels. Since water follows the line of steepest descent available to it, the increasingly large channels carry water on declining gradients in most landscapes, as well. (In detail, stream gradients depend in considerable degree on pre-existing topography, especially in headwaters, and may vary somewhat irregularly.) Fig. 1 presents these circumstances graphically and illustrates some features of the consequent distribution of sediments along drainage systems, along with some derivative properties of the system. The division of the drainage basin described here was first advanced by S.A. Schumm (1977), but not tied so closely to sedimentary mechanics as will be attempted here.

**Fig. 1** Schematic diagram of a drainage basin and some of the principal, systematically varying quantities related to sediment occurrence in stream channels. (a) drainage basin map, illustrating three principal zones of distinctive sediment behaviour. (The drainage network in the upland is not completely drawn.) (b) General pattern of variation along the drainage system of principal quantities related to sediment occurrence and sediment transport (partly after Schumm, 1977). The scale of drainage basin area is approximate and intended only to establish a sense of scale. It will characteristically vary regionally, according to conditions of water and sediment supply. Along a particular channel, topography may also affect the depicted parameters. Distance along the stream channel would increase approximately as the square root of the area scale. (c) General pattern of variation of some quantities of ecological interest that are systematically related to the structure of the drainage system.

One should first notice that the majority of the drainage area is an upland region in which drainage forms and from which sediment is evacuated. In this region, stream channels are directly 'coupled' to adjacent hillside slopes (see Fig. 2) in the sense that sediment mobilised on those slopes may move directly into the stream channel at the slope base. The main valleys constitute something of a 'backbone' for the drainage system. Here, on much lower gradients, sediments mobilised in upland channels may be deposited along the channel, forming an alluvial (= stream transported) substrate and a floodplain. The volume of sediment stored in the system increases dramatically where upland streams first flow into main valley bottoms. These sediments isolate the channel from adjacent slopes, and so the stream becomes 'uncoupled' from hillslope sediment sources. Sediment recruitment and onward transfer become purely consequences of erosion of the streambed and banks. Toward the distal end of the drainage system, where the valleys may open out and become quite broad, sediments may be deposited and stored for long periods of time in floodplains, or in alluvial fans or deltas at the system end.

Because the proximal valley zone both receives sediments directly from upland source areas, and is




subject to remobilisation and onward transfer of sediment by the still energetic rivers, it is exceedingly sensitive to fluctuations in sediment supply from the uplands. Periods of aggradation (excess supply) and degradation (low supply) follow each other according to the development of land surface condition and sediment delivery to upland channels. This zone also represents the upstream limit of fish penetration for most species. Survival and population success for the species that come here are bound to depend in a sensitive way on land surface condition. Hence, human activity in the upland can have a major impact on the condition of aquatic ecosystems in this part of the river system.

The transitions amongst 'upland', 'upland valley' and distal or 'floodplain valley' incorporate the most important thresholds in the riverine landscape. Between upland and upland valleys, the system changes from being dominantly a sediment-evacuating system, to being a sediment accumulating system. Furthermore, the most fundamental threshold of all, the hillslope-channel transition, changes from being dominantly coupled to being dominantly uncoupled. Downstream from this transition, sediment delivery to the channel is primarily by fluvial sediment transport and regularities based on fluvial sediment sorting become prominent features of channel morphology. In upland channels, sediment does not, in general, exhibit systematic trends in texture or sorting because of more or less continuous recruitment from adjacent hillsides (Rice & Church, 1996) but, beyond the upland limit, sediments exhibit consistent trends of fluvial sorting between successive tributary junctions (Sternberg, 1875; Knighton, 1980; Rice & Church, 1998). The downstream transition to 'floodplain valley' marks the onset of persistent and long-term accumulation of sediments and of additional modes of sedimentation. In the field, these transitions may not be abrupt. Furthermore, they may shift position in the system in response to persistent trends of sedimentation brought on by tectonics, climate change, or human activity.

In addition to stored sediment quantity, sediment grain size varies systematically through the drainage system. A wide range of sediment sizes is introduced to the stream channel from upland hillslopes. The stream is not capable of moving the largest, bouldery, material, so it is left as 'lag' deposits along the upland channels. Smaller material moves downstream. In general, the largest size that can be moved is proportional to the force that can be exerted on the streambed by the flow. A modification of Lane's (1955) relation quantifies this proportionality. Statement (1) does not represent a complete correlation, as it is not dimensionally balanced. Replacement of grain size by

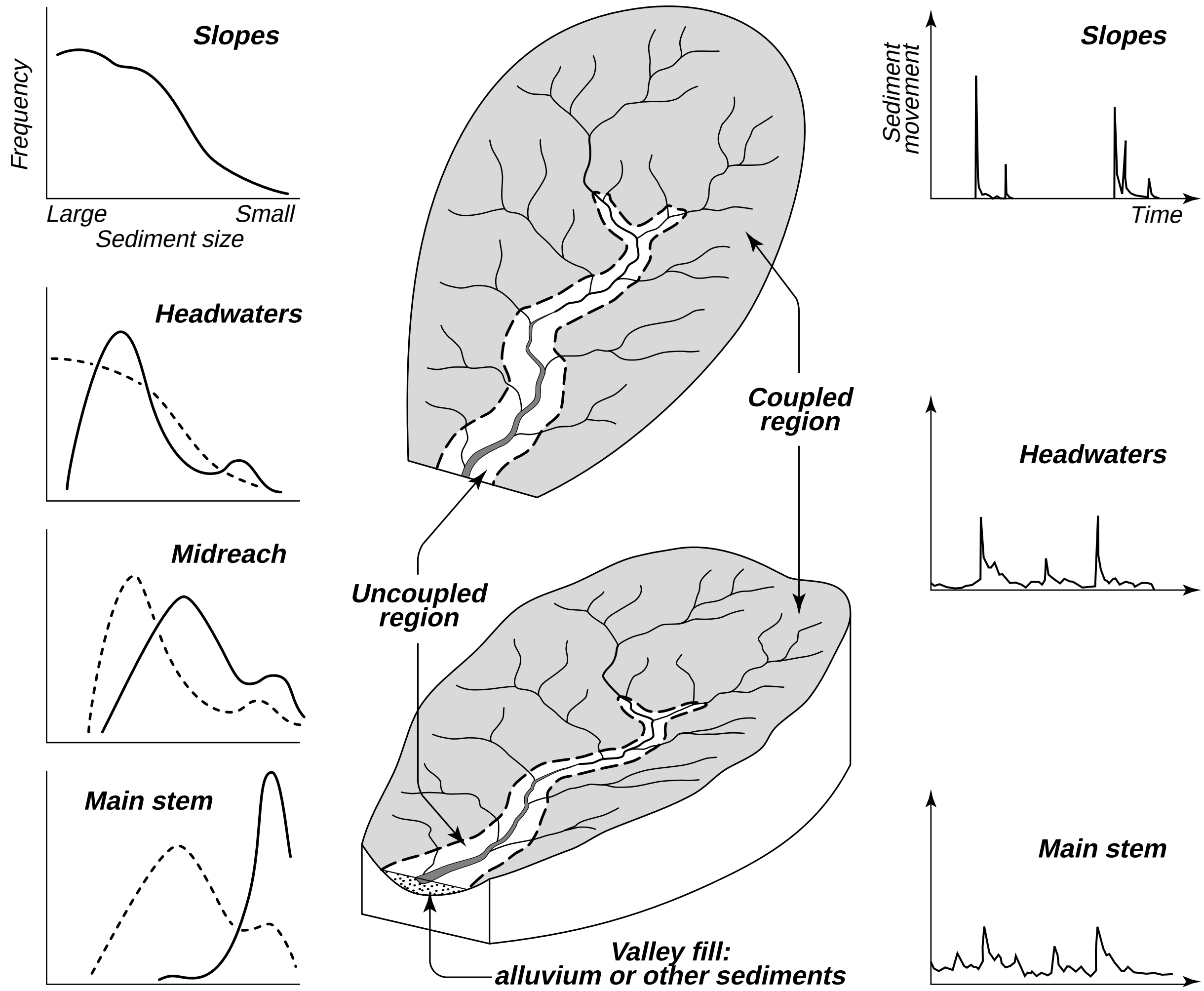



**Fig. 2** Map and diagrammatic schematic views of a drainage basin to illustrate the concept of 'coupling' between a stream channel and adjacent hillside slopes. Near the upstream limit of the decoupled reach there will usually be a significant 'partially coupled' reach, where stream channels move against, and then away from adjacent hillslopes. On the left side of the diagram are schematic graphs of characteristic grain size distributions through the channel system. In each graph, the next upstream distribution is shown (dashed line) so the intervening modification by stream sorting processes may be directly appraised. On the right hand side of the diagram are graphs to illustrate the attenuation of sediment movement down the system. Attenuation is the consequence of increasing mobility of finer material farther downstream, tributary confluences with variations in runoff timing, and of diffusive processes associated with channel flow.

relative roughness, $D/d$, in which $d$ is flow depth, yields a dimensionally rational statement

$$Q_s/Q \sim dS/D \qquad (2)$$

which states that the concentration of sediment in the flow is directly proportional to the shear stress exerted by the flow on the bed and inversely proportional to sediment calibre. For steady, uniform flow (an approximation often employed in studies of sediment transport and river channel form), shear stress is properly expressed as

$$\tau = \rho g d S \qquad (3)$$

wherein $\tau$ is the shear stress (force per unit area; newtons $m^{-2}$) exerted on the stream boundary, $\rho$ is fluid density and g is the acceleration of gravity. Relation (2) is properly written as

$$Q_s/Q = f[\rho g d S / g(\rho_s - \rho)D] \qquad (2a)$$

wherein $g(\rho_s - \rho)$ represents the submerged weight of the sediment and $f$ indicates a functional relation. This relation is equivalent to A. Shields (1936) relation for sediment mobility. Today, it is customarily represented as the limit condition to entrain sediment of a given size, that is, as a limit ratio between the entraining force of the flow and the inertia of the sediment as $Q_s/Q \to 0$. Shields established the competence limit experimentally. In recent years, it has been recognised that the limit cannot be established physically (because of the residual probability for a grain eventually to move at almost any flow, given a sufficiently large streambed area to sample) and so it is now usual to evaluate Shields' limit at some small, finite value of $Q_s$ (see Parker, Klingeman & McLean, 1982; Wilcock & Southard, 1988; Wilcock, 1992).

Downstream, the reduction in S dominates the right-hand side of eqn 2 (see Fig. 1), so the competence (ability of the stream to move sediment of a given size) declines and characteristic sediment sizes become finer downstream. Stream channels are, accordingly, sediment sorting machines. A wide range of grain sizes may be introduced into an upland channel (see Fig. 2). As we move downstream, however, limited stream competence truncates the upper end of the distribution, whilst the ability of the flowing water to suspend and quickly evacuate finer material truncates the lower end. A limited range of grain sizes is left as resident bed material at a particular place in the channel. Where that range is gravel, there is usually a subordinate mode of fine sediment which is wash material caught in the interstices of the bed material deposit. Downstream, the relative importance of the gravel and fine modes varies and the modes move closer together as the coarsest material is left behind, but gravel usually remains dominant to the limit of its occurrence. In distal channels, a single, very well sorted mode in the sand range constitutes the bed material.

Hence, bed material texture varies along a stream channel according to material supply and the range of flows available to move and sort the material.

### *Upland channels*

Channels that are sufficiently steep ought not to accumulate sediment. The limit condition for sediment accumulation can be investigated via the Shields criterion for clastic particle stability in a water flow, which is conventionally specified as

$$g\rho dS/g(\rho_s - \rho)D = \theta_c \tag{4}$$

wherein $\theta_c$ is the limit value of the Shields number (the ratio between fluid stress and grain inertia) for entrainment of sediment of calibre $D$. Rearranging the equation and substituting values for the physical constants, including $\rho_s = 2650$ kg m$^{-3}$ (4) becomes

$$S = 1.65\theta_c(D/d) \tag{4A}$$

In steep channels, the largest clasts are customarily exposed above the water surface and typically have diameter similar to the depth of the channel. Suppose that $D \sim d$ at high flow (i.e. relative roughness, $D/d \sim 1.0$). For individual, wellexposed gravel, cobble or bouldersize clasts, a conventional value $\theta_c = 0.03$ has come to be accepted whereas, for widely graded gravel mixtures, $\theta_c = 0.045$ (cf. $\theta_c = 0.047$ by Meyer-Peter & Muller, 1948), in comparison with Shields' original specification of $\theta_c = 0.06$ for narrowly graded sediment beds coarser than sands. For $\theta_c = 0.045$, $S_c = 0.074$, or 4.2°. At $\theta_c = 0.06$, $S_c = 0.099$ (5.7°). But sediment-choked channels plainly exist on much steeper gradients. In such channels, the stability of the sediments depends not just on particle inertia (weight). It depends on locking arrangements amongst adjacent stones that substantially increase the force necessary to dislodge them.

When $D/d \sim 1.0$ and the width of the channel is less than an order of magnitude greater than the diameter of the largest stones within it, key stones form stone lines that define steps (see Chin, 1989, 1999 for reviews). Pools are ponded behind the steps and the channel forms a step-pool cascade (Fig. 3a). Many authors have attempted to demonstrate the regularity of occurrence of the steps, but they do so by examining averages of step spacings. Average spacings must bear some consistent relation to overall gradient and characteristic keystone size. However, if individual spacings are considered, it becomes evident that steps are essentially randomly placed, no doubt the consequence of the random delivery of the keystones to the channel (Zimmermann & Church, 2001). There are few data on the upper limit gradient of sediment retaining channels. Montgomery *et al*. (1996) studied the analogous case of channels in which sediment was retained by channel-spanning logjams and observed 'forced alluvial reaches' (i.e. reaches in which sediments were retained behind jams) on gradients up to 0.3 (17°). The results exhibited dependence

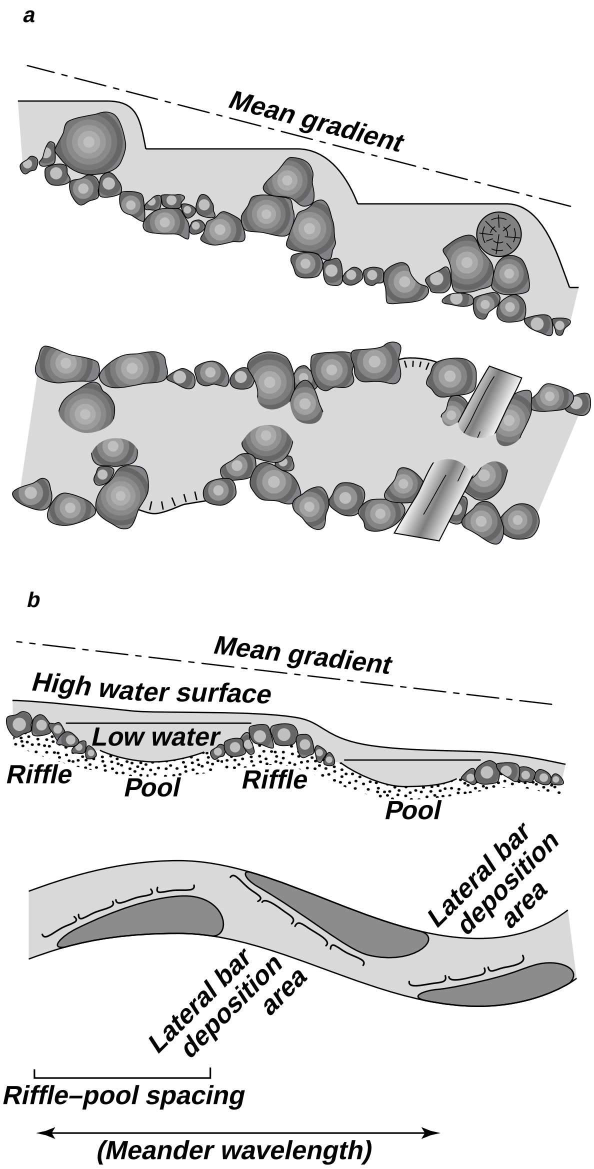



**Fig. 3** Channel elements in high gradient channels (a) step-pool system; (b) pool-riffle-bar system.

upon drainage basin scale, with lower limit gradients observed in larger basins.

On gradients between 4.2° and 2.8° or less (depending upon relative roughness), individual unconstrained grains ($\theta_c \approx 0.03$) remain unstable. Such grains are easily moved, once submerged, unless they are incorporated into imbricate or locked structures. Such structures take the form of stone lines which are more usually non-channel spanning, particularly as channels become larger and resident clasts smaller. At gradients down to 1°, these features develop to take the form of non-channel spanning steps or usually submerged 'stone nets' (Church, Hassan & Walcott, 1998). Relative roughness usually is in the range $0.3 < D/d < 1.0$, so that water flows over the structures remain jet-like or wake-dominated. The overall morphology can be characterised as a 'rapid' (cf. 'plane bed' of Montgomery & Buffington, 1997). The keystones are the largest and least mobile stones of all, typically ones around $D_{99}$. Related features include stone clusters (cf. Brayshaw, 1984) and transverse ribs (Koster, 1978; but they are not, at least in the general case described here, relict antidunes: the features form at subcritical Froude numbers).

Within the range of gradients for rapids to occur, major, discrete riffles and skeletal bar structures begin to appear. This represents the transition to the riffle–pool morphology characteristic of gravel-bed streams on lower gradients in the 'upland valley'. Here, substantial storage of episodically mobile bed material may occur in pool bottoms and on bars superimposed upon the riffles.

Various authors have described the channel morphologies detailed above, and a variety of terminologies and limit criteria have been forwarded. Table 1 summarises some of these. Prior authors have derived their gradient limits from observations and have not, on the whole, referred to sediment stability criteria. The transitions described here appear to be a parsimonious set referred to sediment stability thresholds. There is some evidence that limit gradients are further modified by the scale of the flow. In very small channels the transitions may occur on even higher gradients than those detailed above (Montgomery *et al*., 1996; Halwas & Church in press).

Structural strengthening of the bed increases the critical shear stress to move sediments. Experimental evidence (Church *et al*., 1998) suggests that, in rapids, $\theta_c \approx 0.07$ or more, while field evidence indicates that, in step-pool cascades, $\theta_c > 0.1$. Normal sediment transport in this latter class of channels is restricted to considerably finer material than that incorporated into the key structures (cf. Adenlof & Wohl, 1994). Accordingly, the channels are not strictly 'alluvial' in character. Channel reorganisation is rare and, in cascade channels, at least, is practically restricted to debris flow (mass flow) events. In channels dominated by rapid structures, the larger, structure-forming clasts

**Table 1** Summary on slope and morphology of steep, clastic channels

| Reference | Pool | Glide | Riffle | Rapid | Cascade | Comments |
|---|---|---|---|---|---|---|
| Bisson *et al*. (1981) | | | $S < 0.04$ | $S > 0.04$ | Steepest rock, boulder, or LWD jams | Many channel sub-units identified |
| Sullivan (1986) | $S < 0.01$<br>Dammed, 0.002<br>Scour, 0.006<br>Plunge, 0.009 | $0.01 < S < 0.02$ | $0.01 < S < 0.04$<br>$\langle S \rangle = 0.022$ | $S > 0.04$<br>$\langle S \rangle = 0.07$ | $0.04 < S < 0.16$<br>$\langle S \rangle = 0.068$<br>(a) step-pool<br>(b) slip-face | Channel units are 4–8 w in length |
| Grant, Swanson & Wolman (1990) | $S \sim 0.005$<br>Tranquil flow | – | $\langle S \rangle = 0.011$<br>not ribbed | $\langle S \rangle = 0.029$<br>Ribbed | $\langle S \rangle = 0.055$<br>Boulder;<br>bedrock | Channel subunits 0.4–0.8 w;<br>channel units 1–10 w;<br>channel reach 100–1000 w,<br>steps have $\langle S \rangle = 0.17$ |
| Wood-Smith & Buffington (1996) | Closed topographic depression<br>(a) obstructed<br>(b) not obstructed | $S_b < 0.02$ | $0.02 < S_b < 0.04$ | – | $S_b > 0.04$ | Modified from Bisson *et al*. (1981) and Sullivan (1986); overall stream gradients < 0.025 |
| Montgomery & Buffington (1997) | *$\langle S \rangle = 0.012$ for pool/riffle unit*<br>*$S < 0.031$; $<0.036$ forced by LWD* | *''Plane bed''*<br>*$\langle S \rangle = 0.023$*<br>*$0.0015 < S < 0.04$* | – | *''Step-pool''*<br>*$\langle S \rangle = 0.044$*<br>*$0.02 < S < 0.076$* | *$\langle S \rangle = 0.11$*<br>*$0.020 < S < 0.20$* | *Gradients are means over several channel units* |
| Halwas & Church (in review) | *Closed topograph. depression*<br>*(a) dammed*<br>*(b) plunge* | *$\langle S_b \rangle = 0.06$* | *$\langle S_b \rangle = 0.09$* | *step-pools*<br>*$\langle S_b \rangle = 0.20$* | *$\langle S_b \rangle \sim 0.45$ (boulders)*<br>*$\langle S_b \rangle \sim 0.49$ (rock)* | *Small channels, $S_b < 0.25$, generally* |
| Summary | Closed depression | $S < 0.02$ | $S < 0.04$ | $S < 0.10$;<br>$D/d < 1.0$ | $S > 0.04$;<br>$D/d > 1.0$ | |

$S$ = water surface slope; $S_b$ = bed slope. Most results follow Bisson *et al*. (1981) and Sullivan (1986), with amendments by Grant *et al*. The results of Montgomery & Buffington (1997); and of Halwas & Church (in press), are not comparable with the rest (*in italics*). Montgomery & Buffington (1997) measured mean gradients over some distance; Halwas & Church (in press) observed much smaller streams than all others.

are moved during major floods (recurring perhaps once every few years), so that the condition of 'partial transport' (Wilcock & McArdell, 1997) characterises the larger material found in these channels.

### *Valley (alluvial) channels*

Once relative roughness declines below about 0.3, the possibility exists for the stream to begin to accumulate vertically stacked sediments. This happens once grain size declines and depth increases sufficiently. Then channels in gravel develop a well-expressed pool-riffle-bar system (Fig. 3b) which has affinities with meanderform development. Such channels have been extensively studied (e.g. Richards, 1976; Lisle, 1982; Madej, 1999). The main structural feature of the bed is imbrication, although features characteristic of rapids may still be seen on riffles. For imbricated gravel beds there is a wide consensus that $\theta_c \approx 0.045$ (Wilcock & Southard, 1988), whence, for $D/d \leq 0.3$, $S \leq 0.022$ (1.3°). Various authors report riffles on gradients up to 0.04 (2.9°). Primary pools (those dammed upstream of channel-spanning riffles: see Fig. 3b) recur with approximately regular spacing of 5–7 channel widths, but a wide variety of more local scour pools exists as the result of flow convergence set up by obstructions (boulders, large woody debris, bank projections: see Lisle, 1986) that play a significant role as habitat units (Bisson *et al.*, 1981). The primary riffles may remain anchored in place, or may migrate slowly along the system according to the relative mobility of the material forming the streambed and the overall tortuousness of the channel. Abrupt bends may anchor otherwise migratory sediment accumulations.

In company with steeper channels, sediment transport through pool-riffle channels in gravel remains limited by stream competence and sediment structures, so that 'partial transport' remains normal. Again, this implies the occurrence of patches of substrate that may remain stable for substantial periods. This feature, together with the characteristically high permeability of gravels, supports invertebrate communities that extend into the subsurface. It is conjectured that the stable patches provide source areas for benthic recolonisation of adjacent areas scoured during high flows (e.g. Matthaei *et al.*, 2000 and references therein).

The transition from gravel to sandy substrates represents another signal threshold in the channel system. This transition is almost always accompanied by a discontinuity in the downstream trend of grain size. Despite long study, the phenomenon remains imperfectly understood (see Sambrook-Smith & Ferguson, 1995). A fundamental distinction of sand-bed channels is that, because of the small size and low inertia of individual grains, bed material is mobile over a wide range of flows and substantially the entire bed takes part in the sediment exchange with the flow. The bed becomes inherently unstable and exhibits a range of wave instabilities which take the form of sand ripples, dunes and longer period sand waves. The frequently mobile bed forms a much less hospitable habitat than do gravel substrates.

Stream channel stability now depends not on bed stability but on bank stability and gross sediment exchange. This refocuses our attention away from thresholds based on stream competence toward ones based on stream power.

## Sediment transport and channel state

The capacity of a stream to transport sediment has been analysed in terms both of tractive force and of stream power. Insofar as power implies the capacity to do work (in this case, sediment transporting work), the latter concept is attractive (Bagnold, 1966). It is also convenient for the consideration of related morphological transitions in stream channels.

The two factors, flow magnitude and gradient, combine to determine the power of the stream, as indicated in statement (1a) above. Re-expressing that relation in terms of stream power per unit width of the channel, $\omega = \rho g Q S / w$ (i.e. dividing through by channel width) and reintroducing the scaling value, $d$, as in eqn 2, we obtain

$$q_s(x) \sim \omega(x) \cdot d(x) / D(x)$$

i.e.

$$q_s = f[\omega \cdot d/D] \tag{5}$$

wherein $x$ is a dummy variable of position (which is dropped in eqn 5) to signal that the indicated variates change significantly with distance from headwaters. This is Bagnold's (1980) formula for bed material transport. Introducing $g^{1/2}D^{3/2}$ as a scale [a 'volumetric Einstein scale', following scaling arguments introduced by Einstein (1950)] and defining scaled volumetric transport $\xi = q_s / g^{1/2}D^{3/2}$, then

$$\xi = f[(\omega/g^{1/2}D^{3/2}) \cdot d/D] \quad (5a)$$

From empirical investigations (Martin & Church, 2000), we find

$$\xi = f[(\omega/g^{1/2}D^{3/2})^{3/2} \cdot (d/D)^{-1}] \quad (5b)$$

or

$$q_s \sim (\omega/g^{1/6}D^{1/2})^{3/2} \cdot (D/d) \quad (5c)$$

As streamflow and gradient generally vary inversely with each other, stream power is relatively conservative along the drainage system (cf. Fig. 1: see observations in Knighton, 1999). Though stream power may denote the potential for the stream to do work, as in moving sediment, the work actually accomplished also depends upon the efficiency with which energy can be transmitted from the flowing water to the sediment. In most channels, that efficiency is very low. Hence, the numerical coefficient that remains to be determined in eqn 5c is of order $10^{-2}$.

In order for sediment to be transferred in an orderly fashion through a stream channel, stream power must tend to increase with channel scale (to preserve the sediment transporting capacity). For a given channel width, ω would tend toward a constant value and the transport *intensity* (transport per unit channel width, or 'specific transport', $q_s$) would remain constant. It is well known that channel scale is set by $w \propto Q^{1/2}$. Hence, to maintain ω constant, $S \propto 1/Q^{1/2}$ (from the definition of ω). This relation is interesting, for it turns out to discriminate channels of different planform morphology, or different 'style' (Fig. 4), leading to distinctive riverine landscapes. The diagram shown in Fig. 4 was introduced by Lane (1957), but the most famous form of it is due to Leopold & Wolman (1957). The diagram has been modified by many subsequent authors and the underlying relation has been studied in an increasingly sophisticated way (e.g. Parker, 1976). The relation can be proposed directly from relation (1a), which ties it directly to sediment transport intensity, but that relation also implies a dependence on sediment grain size, $D$. Indeed, there is such an effect, first demonstrated by Henderson (1961), although it is of second order (and must be an incomplete correlation, as relation (1a) is not a rational statement).

There is not a strict physical interpretation of the channel pattern discrimination afforded by Fig. 4. Most reasonably, however, it signifies the relation between channel style and the quantity of bed material supplied to the channel. If loci $SQ^{1/2}$ denote constant transport intensity, then a supply larger than the indicated quantity would cause aggradation and a smaller supply would cause degradation or armouring (i.e. structural modification) of the channel bed. A cumulative history of such episodes would establish some relation between the quantity of potentially mobile sediment stored in the streambed and the current rate of transport.

When bed material supply exceeds transporting capacity, sediment is deposited in the channel bed. To maintain conveyance, the water must find room to go around these deposits. In relatively coarse-grained materials (sands and larger), this is accomplished, in sufficiently large channels, by lateral erosion – by

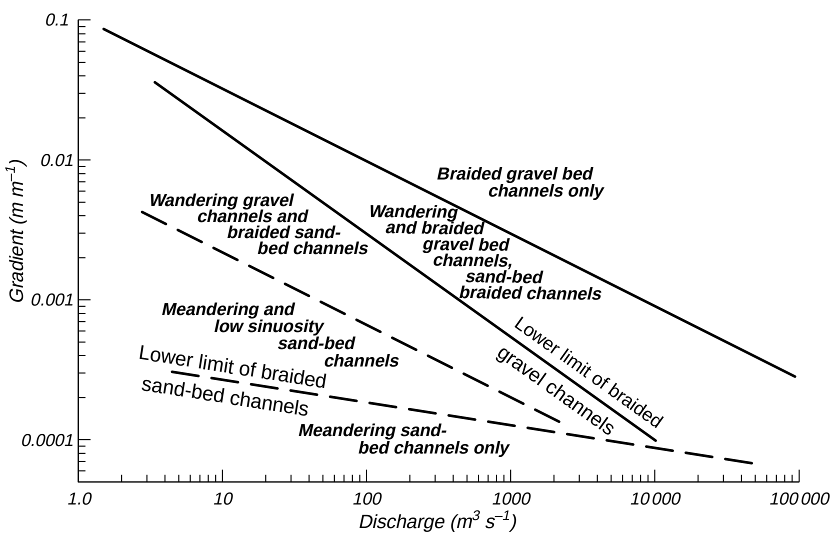



**Fig. 4** Fields of river channel morphological pattern within the domain of $S$ versus $Q$. $Q$ is the 'channel-forming discharge', which is variously interpreted to be bankfull flow or mean annual flood. [partly after Leopold & Wolman (1957)] The different positions of gravel and sand transitions show the effect of sediment calibre, which is not otherwise quantified in this display. Note the distinct plotting positions of 'upward' and 'downward' transitions in channel morphology, the consequence of varying bank constitution and bank strength in the different regimes.

erosion of the banks. This process introduces a replacement sediment load into the channel. A 'sufficiently large channel', in this context, is a fully alluvial one. By the definitions in this paper, it is a channel that may be found in an upland valley or distal valley. If this process proceeds far enough, the channel becomes sufficiently wide that it becomes braided; that is, the bed material deposits become widely exposed at moderate and low flows. The river must also steepen to maintain transport intensity in the increasingly wide (and shallow) channel. This effect is achieved either by channel straightening, so that the river flows more directly downvalley, or by persistent aggradation toward the head of the reach. This shifts the channel upward in the $S$ versus $Q$ domain. Discriminant lines of channel pattern in this diagram represent important thresholds as characteristically different morphologies and habitat features are delimited by these transitions. Channels in the sensitive transition region from upland to upland valley are particularly prone to pattern shifts as the result of changing sediment supply from upland tributaries, often aggravated by human activity both in the upland and along the channel.

Channels falling in the braided region create riverine landscapes dominated by broad gravel or sand flats and multiple shallow channels. During low flow many channels dry up. The adjacent floodplain is low and often laced with recently abandoned channels, or with flood channels. It is often turf or shrub dominated, depending upon the regional environment but also upon the recent history of channel shifting, which may be abrupt. On balance, these environments are hostile for aquatic life. Cover is low, there is comparatively little input of nutrients, and risk of stranding is high. Extremes of water temperature may be experienced. Nevertheless, important habitats may exist within the channel zone and in the floodplain. Significant flow occurs below the surface in relatively coarse fluvial deposits and groundwater resurgence maintains spring-fed channels within the channel zone during periods of moderate to low flow. In the floodplain, minor channels (including tributary streams) may offer reliable flows, edge cover and drop-in food sources. Relatively rapid groundwater circulation through coarse sediments maintains flow in these channels, which are often relatively low in comparison with channels in the main, aggraded channel zone. Flooding typically commences by seepage of groundwaters out of the main channel zone during rising flows.

An interesting feature of the diagram of channel morphologies is that, whilst 'upward' transitions from single-thread to multithread channels follow threshold loci of constant $SQ^{1/2}$, the downward transitions are distinct and do not follow such loci. The distinction of thresholds is related to bank strength, which has a strong influence on the transition from single-thread to multithread channels (Millar & Quick, 1993). In simple terms, a single-thread channel with root-reinforced banks will withstand erosional attack more effectively than the nonreinforced banks that are apt to occur within a braided channel zone (Millar, 2000); hence an asymmetric threshold occurs between the two states. But channels in the intermediate region between the two transitional loci remain only conditionally in their current state. Once bank condition is changed, the channel will change into the successor state. The transition from braided to single-thread channel is sometimes induced by deliberate bank hardening to limit river erosion of adjacent land.

The downward thresholds also behave differently in sand and in gravel. The downward threshold occurs at increasingly low values of $SQ^{1/2}$ in larger gravel-bed channels (i.e. at decreasing transport intensity), whereas it occurs at increasingly large values in sand-bed channels (i.e. at increasingly large values of $SQ^{1/2}$). In gravels increasingly large channels are increasingly deep, so that bank erosion is increasingly apt to occur below the effective rooting depth of bank vegetation. Hence, vegetation declines in importance as a bank-reinforcing mechanism as channels grow larger. Gravel-bed channels in the intermediate state take on a distinctive form, combining the features of braided and irregularly sinuous, single-thread channels, and including channel islands. This has been termed a 'wandering' state by Canadian investigators (cf. Neill, 1973; Desloges & Church, 1989). These channels create a riverine ecosystem with an extended channel zone and adjacent floodplain distinguished by many side- and back-channels. The channel type provides an excellent habitat, especially for salmonine fishes, which prefer fast water and gravel substrate. The wide variety of channel form offers a full range of depth and velocity conditions over most of the flow range; the channel islands offer a high ratio of bank to total channel length and abundant inputs of organic matter and cover (Gregory et al. 1991; Nakano & Murakami, 2001). Vigorous groundwater circulation through gravels maintains substantial hyporheic communities (e.g. Malard *et al.* 2002). Seasonal flooding provides further important connections between the channel zone and the adjacent floodplain. Again, human intervention in the floodplain often restricts the channel system by cutting off back-channels and even by confining the main channel zone (usually successfully, as the channels are a transitional type anyway).

In sand-bed channels, the 'upward' and 'downward' transitions are affected by the importance of suspension of the increasingly dominant fine material in larger systems. Suspension of a significant proportion of the sediment load changes the river from one that accretes sediment deposits laterally to one that accumulates deposits vertically from overbank floods, creating narrower, deeper channels with substantial bank strength gained from the cohesion of the fine-grained deposits. Sinuous, single-thread channels and a more distinctly terrestrial floodplain are the major elements of the riverine landscape. Here, periodic (often seasonal) flooding provides a major mechanism to promote transfer of nutrients, material and organisms. With more restricted groundwater circulation through finer sediments, the inception of flooding usually occurs by overbank passage of water, initially at low points. This connection has been broken by flood control measures in many places and, more than in most other river types, channelisation to rationalise land access or to protect communications routes has simplified river channels. The sand thresholds still have not been studied in detail and it is likely that considerably more nuanced variations exist in the associated riverine landscapes than can be described here.

An additional major class of channels has not been fully incorporated into the discrimination of channel patterns implied by the foregoing discussion. These are anabranched channels (Nanson & Knighton, 1996) – multiple channels with semipermanent, vegetated islands. Nanson & Knighton recognise six subclasses, of which the last two correspond to wandering gravel bed channels and divided, stable upland channels, respectively. Classes 1 and 2 are silty and sandy, suspension dominated rivers that plot in the single-thread, meandered region of the *S–Q* diagram, but evidently have much higher suspension or mixed suspension and sandy bedloads than can be transferred through a single channel. These appear to be fine-grain equivalents of systems with high sediment storage. These types commonly occur in river deltas and the channel pattern is described as 'anastomosed'. Class 3 is a mixed-load meandering type, again with high sediment storage and channel division. Class 4 features long, ridge-like islands that appear to form initially as streambank gallery forests in semiarid and monsoon tropical environments. This appears to be a rather special circumstance.

Sandy to silty channels may contain highly diverse fauna and aquatic flora, but their productivity depends essentially upon the connectivity between the main channel and backwaters (Amoros & Bornette, 2002). Hence, anabranched channels are particularly important from a biological viewpoint. Backwaters contribute a high proportion of instream primary productivity, they provide escape terrain from the main currents of floods, and they provide a mosaic of habitat niches that shelter diverse species (Ward *et al.*, 2002). Wetlands, including seasonally flooded floodplains, form an important part of such riverine systems. Anabranched river deltas, with extensive interchannel wetlands, may constitute the most productive 'terrestrial' habitats of all. Again, extensive engineering of channels of these types have been undertaken, particularly in Europe, where widespread draining of wetlands occurred in the 18th and 19th centuries to reduce flooding, to increase arable land area, and to combat diseases (see Petts *et al.*, 1989).

The introduction of suspension load in the foregoing paragraphs (or, probably more significantly, wash versus bed material load) introduces a new way of considering channel thresholds in relation to sediment transport and ties together considerations of competence and sediment supply. Fine material commonly is transported at Shields ratios far in excess of critical values. Dade & Friend (1998) have estimated that rivers dominated by suspension of silty or fine sandy sediments have $\theta \sim 10$ in channel-forming flows. Mixed-load channels, with medium to coarse sand or fine gravel bedload appear to have $\theta \sim 1.0$, and bedload dominated channels (gravel-bed channels) have $\theta \sim 0.04$ (i.e. very close to threshold). To retain regime stability (i.e. stability of channel form, but not necessarily absolute stability of position), these various channels must have either dramatically different bank strength, or different gradients. Often, they appear to possess both, and so they plot distinctively in the *S*–*Q* diagram (Fig. 4). Sediment properties, hence the transport mode, create distinctive morphologies, as first pointed out by Schumm (1963). An extensively modified form of Schumm's (1963) channel classification table is presented as Table 2. It provides an elementary division of channel styles and can be interpreted to provide a first-order division of riverine landscapes.

On balance, it appears that the morphological role of sediment storage in stream channels, hence of water and sediment regime changes over considerable periods of time, deserve greater attention than they have received in determining the important thresholds that occur between channel styles, hence in determining channel morphology and the character of riverine landscapes.

## Flow transitions

The discussion to this point has considered channel thresholds created by systematic variations in stream competence and sediment supply. In the short term, most river channels are relatively stable, but transitions still occur because flow changes seasonally and synoptically, so the channel is filled to various degrees. A classical characterisation of these short-term stages is shown in Fig. 5. At low flow, a river is shrunken within its banks, so the channel edge is some distance from the riparian fringe. The effect is apt to be greater in wide, shallow channels – most especially in braided channels – than in other types. An intermediate water level of some importance is the 'bar top' level. Bar tops become the channel edge or are just submerged. Water comes adjacent to or moves into the lower fringe of riparian vegetation. This stage probably presents the widest range of in-channel habitats that occurs, for all depths are present.

At higher flows, the channel fills toward outer bank tops, and bar-edge or bar-top slack water begins to disappear. Bankfull flow may present significant difficulties for many stream biota within the channel as 'escape terrain' may become quite restricted. Once flow is overbank, escape may be outside the channel. The overbank stage is also recognised to be disproportionately important for the exchange of organic matter and nutrients that is afforded between the river channel and adjacent land surface, so much so that the 'flood pulse' concept is the key element in understanding the ecosystem in distal rivers with significant floodplains (e.g. Junk, Bayley & Sparks, 1989; Ward *et al*. 2002).

Whilst each of the stages described above may be recognised to represent passage through important thresholds in habitat character and distribution in a channel and floodplain, it represents far from a complete story about flow transitions. The picture appears to be appropriate for fine-grained, vertically accreting sinuous or anabranched channels (many of which are tropical or subtropical and experience an extended flood season). However, it may not represent so fairly the regime of flow transitions in laterally accreting sinuous, wandering or braided gravel-bed systems, many of which are temperate to boreal and experience synoptic rather than seasonal flooding. Furthermore, it does not represent upland streams at all, many of which are effectively confined within high, sloping banks at all realisable stages.

Gravel-bed rivers have wide, relatively shallow channels and exhibit an irregular, lateral style of sediment accretion and floodplain construction with limited vertical development. Chute channels behind bars, which may become secondary channels behind islands, switching of the main channel between alternate routes around bar-island complexes (in effect, avulsions within the channel zone), and persistence of 'back channels' after the main channel has shifted away, are all characteristic. All these features create large channel-zone conveyance with limited duration of overbank flooding initiated by seepage during storm-period peak flows. Escape terrain is represented by the side and back channels. Important thresholds are represented by bar-top inundation and side-channel activation. The flood pulse takes a quite different form.

In boreal rivers, high water and floodplain inundation may be represented not by high flow at all but by ice jam flooding. In these rivers, a distinctive transition occurs at freeze-up, when the water column may become largely isolated from the atmosphere for a more or less protracted period. In comparison with seasonal flood regimes, these phenomena and their ecological effects have been little studied.

Upland streams may exhibit the most dramatic thresholds of all. Flows at all stages are commonly retained within a confined channel zone. The exaggerated transitions between successive pools and

**Table 2** An elementary classification of alluvial river channels and riverine landscapes

| Characteristic Shields' number* | Sediment type | Sediment transport regime | Channel morphology and stability | Riverine landscape |
|---|---|---|---|---|
| 0.04+ | Cobble-gravel | Bedload dominated; low total transport in partial transport regime | Cobble-gravel channel bed; single-thread or wandering; highly structured bed; relatively steep; low sinuosity; $w/d > 20$, except in headwater boulder channels; relatively stable | Determined by adjacent topography; limited riparian zone, except in forested valley bottoms, where log jams may induce backchannel development (Hogan, Bird & Hassan *et al.*, 1998) |
| 0.1+ | Sand to cobble-gravel | Bedload dominated, but possibly high suspension load; partial transport to full mobility | Gravel to sandy gravel; single-thread to braided; limited, local structure development; complex bar development by lateral accretion; moderately steep; low sinuosity; $w/d$ very high; subject to avulsion and frequent channel switching; | Wandering to braided channels; possibly extensive side-channel development; low floodplain with sandy top member; mosaic forest development or heath/tundra |
| 1.0± | Sand to fine gravel | Mixed load; high proportion moves in suspension; full mobility with sandy bedforms | Mainly single-thread, irregularly sinuous to meandered; lateral (point) bar development by lateral and vertical accretion; levees present; moderate gradient; sinuosity < 2; $w/d < 40$; mainly progressive lateral instability | Single-thread channels, irregular lateral instability or progressive meanders; sandy floodplain, mainly dry; strip forest development along accreting channel margins |
| Up to 10 | Sandy channel bed, fine sand to silt banks | Suspension dominated with sandy bedforms | Single-thread, meandered, limited point bar development; vertical accretion in the floodplain; levees prominent; low gradient; sinuosity >1.5; $w/d < 20$; serpentine meandering and cutoffs | Single-thread, highly sinuous channel; loop extension and cutoffs; fine sand to silty floodplain has extensive wetlands away from the channel; strip forest development along accreting channel margins; water table dominated forest succession elsewhere |
| <10 | Silt to sandy channel bed, silty to clay-silt banks | Suspension dominated; minor bedform development | Single-thread or anastomosed channels; vertical accretion on the floodplain; levees prominent; very low gradient; sinuosity > 1.5; $w/d$ <15 in individual channels; stable or slowly shifting channels | Single-thread or anastomosed channels; common in deltas and inland sedimentary basins; extensive wetlands and floodplain lakes; vegetation succession dominated by water table levels in terrestrial or semi-lacustrine environment |

*Modified after Dade & Friend (1998). Values apply to channel-forming (i.e. flood) flows.

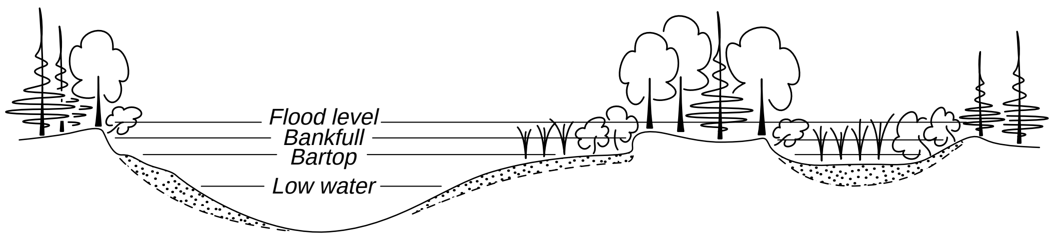



**Fig. 5** Cross-section of a floodplain river, showing significant water level thresholds.

riffles, rapids or steps that occur at low and intermediate flows become drowned, at least to surface appearance, at high flow, when the entire channel may become a rapid torrent. The large clastic or wood elements that constitute the principal material of the channel boundary may, however, still afford significant refuge from the main flow. The high stability of the principal elements in these channels represents a vital tradeoff against the lack of side-channel or off-channel escape terrain. At high flow, important thresholds occur in these channels between the high velocity main current and the eddying backwaters and interelement openings in the bed and banks.

The incidence of flow events that reform the channel by accomplishing significant cumulative bed material transport varies systematically through the flow regime types described. The classic description of the geomorphological effectiveness of river flows was given by Wolman & Miller (1960), but they described only rivers of intermediate to large size. In upland areas, the most significant floods derive from highly localised extreme rain. Whilst such events may be regionally common, they rarely impinge on any one place. The result is that channel-reforming floods – which may have to move very large material – are relatively rare in an individual channel (see Fig. 2). Return periods of once in 30–50 years are not uncommon. Downstream, flow convergence occurs as tributaries join the main channel, so that extreme runoff from somewhere in the upstream system becomes more common. So does, however, attenuation of the flood wave and of the forces necessary to move the materials (themselves smaller) that make up the channel boundary. The result is a steady reduction in the return period of channel reforming events. At the downstream limit of large, sand-bed channels, material is moving nearly continuously. The violence of an extreme headwater flood, which may clear much of the biota right out of the channel is, then, succeeded by a long recolonisation period. Downstream, such system-resetting events are rare or absent, and there are much greater opportunities, both functionally and spatially, for biota to adapt to the synoptic or seasonal stresses presented by the flow regime. Thresholds of water level or flow imply quite different things, then, at different places in the system.

## Summary

In this paper, I have endeavoured to consider thresholds in terms of the transitions that give distinctive form or introduce distinctive processes into the river. The main points are as follows.

1. The main controlling factors are the flow regime and the regime of sediment supply. Additional important factors include topographically determined gradient, riparian vegetation (hence bank strength) and, in northern rivers, seasonal ice regime. Sediment supply confers the greatest distinctiveness on river morphology at all scales, but with growth in the scale of the system the range of sediment regimes becomes steadily more restricted toward sandy-silty types.
2. Flow competence plays a key role in determining river morphology in headwater and upland valley gravel-bed rivers. Downstream, sand and finer sediment may be mobilised over a wide range of flows, so that competence becomes less significant. The balance of sediment supply and transport capability – how far the river exceeds the competence threshold – influences the bed structure and the style or pattern of the channel, creating distinctive physical habitats in different parts of the river. Moving along the channel system, a sequence of thresholds is established by significant changes in the processes by which sediment is moved and stored. These also determine the nature of the adjacent floodplain, whether laterally or vertically constructed, whether normally dry or semi-aquatic. Altogether, these features present

distinctive riverine landscapes and distinct habitat possibilities at different places in the system. Under the influence of sediment supply and transport capability, rivers may change their character in the long term. For relatively large rivers, the long-term may exceed $10^3$ years (Dade & Friend, 1998; see also Church, 1995).

3. Streamflow changes according to transient weather and seasonally and thereby establishes a sequence of temporal threshold crossings in channel environments. In alluvial channels, the interaction between water level, water currents, and morphology creates a series of thresholds related to the interaction between the channel and the adjacent riparian zone. These interactions take on distinctive forms in seasonally and synoptically dominated flow regimes, and between fine-grained (vertically building) and coarse-grained (laterally accreting) systems. In gravel systems, groundwater levels under the floodplain play an important role in mediating ecological processes in the floodplain whereas, in fine-grained systems, surface flooding is a dominant control. Upland systems are again distinctive, with flow-related changes being largely restricted to the channel. Changes throughout the system in the frequency of recurrence of extreme ('system-resetting' or channel reforming) events play an important role in the coping style of aquatic creatures and riverine ecosystems.

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