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Bedforms, Structures, Patches, and Sediment Supply in Gravel-Bed Rivers

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16.1 Introduction

In sand-bedded rivers there is a well-known sequence of bedforms that develops with increasing flow strength. For a specific grain size, the sequence includes: lower stage plane bed, ripples, dunes, upper stage plane bed, antidunes, chutes-and-pools, and cyclic steps (see recent reviews in Venditti (2013) and Cartigny *et al.* (2014)). These features control flow resistance and bed-material sediment flux rates, and leave sedimentary structures in the rock record that are often used to infer paleoflow conditions from an understanding of the modern sedimentary processes responsible for bedform development (Venditti 2013). In gravel-bedded rivers, a comparable sequence of bedforms with increasing flow strength has not been identified. Nevertheless, these rivers develop a distinct suite of bed features that are larger than individual clasts and smaller than channel-scale features (e.g. bars or step-pool features). These bed features include (i) gravel dunes (e.g. Carling 1999), (ii) mobile, migrating patches typically referred to as bedload sheets (e.g. Whiting *et al.* 1988; Nelson *et al.* 2009), (iii) sand bedforms developed over an otherwise immobile gravel bed (sand ribbons, barchans and dunes; e.g. Kleinhans *et al.* 2002), (iv) pebble clusters (e.g. Dal Cin 1968; Brayshaw 1984), (v) stone lines (Larrone and Carson 1976) or transverse ribs (e.g. Koster 1978; Allen 1984), and (vi) reticulate stone cells (e.g. Church, Hassan, and Wolcott 1998; Hassan and Church 2000). There have been many important insights into how these bed features develop and how they influence flow dynamics and sediment transport rates, but there has been little insight into how or whether the features are inter-related, and the conditions required for the emergence of one type of bed feature as opposed to another. Here, we demonstrate the critical role that sediment supply and differential mobility of the grain-size distribution play in the types of bedforms that emerge. We end with a discussion about what controls the emergence of these various bedforms, bed-surface structures, and patches in a common conceptual framework.

16.2 Bedload Transport, Sediment Supply, and Bed Mobility

It is commonly assumed that the primary control on bedload transport rates is the flow strength, often characterized by the mean boundary shear stress. The idea has been embedded in a wide range

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of bedload transport formulae (e.g. Meyer-Peter and Muller 1948; Parker, Klingman, and McLean 1982; Parker 1990). However, observations in rivers and experimental channels have revealed that bedload transport rates can vary by several orders of magnitude under steady uniform flow conditions due in part to the emergence of various bedforms, patches, and structures of an armored gravel surface. Nelson *et al.* (2009) provide an example of the interplay between bedload transport and patches of similarly sized bed material (Figure 16.1). They fed a unimodal, lognormally distributed gravel with a median size D_{50} = 8 mm and range from 2 to 32 mm to a 25 m long and 0.86 m wide channel. Flow rate in the channel was held constant and was above the threshold to move the gravel mixture. When the sediment output rate matched the supply rate, distinct patches were identified and mapped and the supply to the channel was reduced.

The patchiness that emerged on the bed in the experiments is one type of a wide variety of sediment patches of similar grain size and sorting that occur in gravel-bedded rivers. Following the terminology proposed for bar types (e.g. Seminara 1998), sediment patches can be identified as either "forced" (spatially persistent associated with topographic controls), "fixed" (spatially persistent due to coarsening), or "free" (migrating patches) (Nelson *et al.* 2009). The migrating free patches shown in Figure 16.1b have been referred to as bedload sheets, typically described as low-amplitude bedforms with heights of 1–2 coarse grain diameters that fine toward the tails (Whiting *et al.* 1988). The downstream migration of these patches cause a nearly two orders of magnitude change in the bedload transport rate (Figure 16.1a). Recking *et al.* (2009) document entirely similar behavior in the transport of a tri-modal gravel mixture due to the formation and migration of bedload sheets.

As the sediment supply to a channel is progressively reduced, the bed-surface grain size becomes substantially less heterogeneous (Figure 16.1b). This occurs because the active bedload transport corridor on the bed narrows (Lisle, Iseya, and Ikeda 1993; Nelson *et al.* 2009) and the patches become less distinct (Figure 16.1b). This feeds back on the sediment transport rates, which progressively exhibit less variability (Figure 16.1a). Once the sediment supply to the channel is removed, the bed becomes uniformly coarse (Figure 16.1b). A more subtle response that occurs due to the reduction of sediment supply is a progressive change in the size distribution of the particles transported as bedload. Under conditions of high sediment supply, the channel moves particles in proportion to their presence in the sediment supply, but as the supply is reduced, some of the coarser particles get left behind in the bed and coarse patches expand. Below, we explore how sediment supply and differential mobility of the grain size distribution impact the emergence of different types of bed features in gravel-bedded rivers. But first, we provide some quantitative measures of the sediment supply and differential mobility.

16.2.1 Sediment Supply

The rate at which bedload material is supplied to a channel exerts a strong control on the bed-surface grain size (Kuhnle and Southard 1988; Dietrich *et al.* 1989; Buffington and Montgomery 1999; Nelson *et al.* 2009). In channels where sediment supply is not limited, there is no difference between the surface and subsurface grain-size distributions (Laronne and Reid 1993; Laronne *et al.* 1994; Reid and Laronne 1995). This is commonly observed in unvegetated arid watersheds where there is ample supply of all sediment sizes and flows are typically flashfloods with a rapid decline in flow that prevents reworking of the surface after general bedload movement has ceased (Laronne *et al.* 1994). In contrast, seasonal and perennial streams can exhaust their sediment supply, forcing adjustments to the bed surface (Reid and Laronne 1995).

Channels where the sediment supply is limited with respect to the transport capacity are typically armored (Parker and Klingeman 1982) and have a bed surface slightly coarser than the bedload, which often has a grain-size distribution similar to the subsurface. When the sediment supply is nearly equivalent to the transport capacity of the channel, active bedload transport takes place across much of the

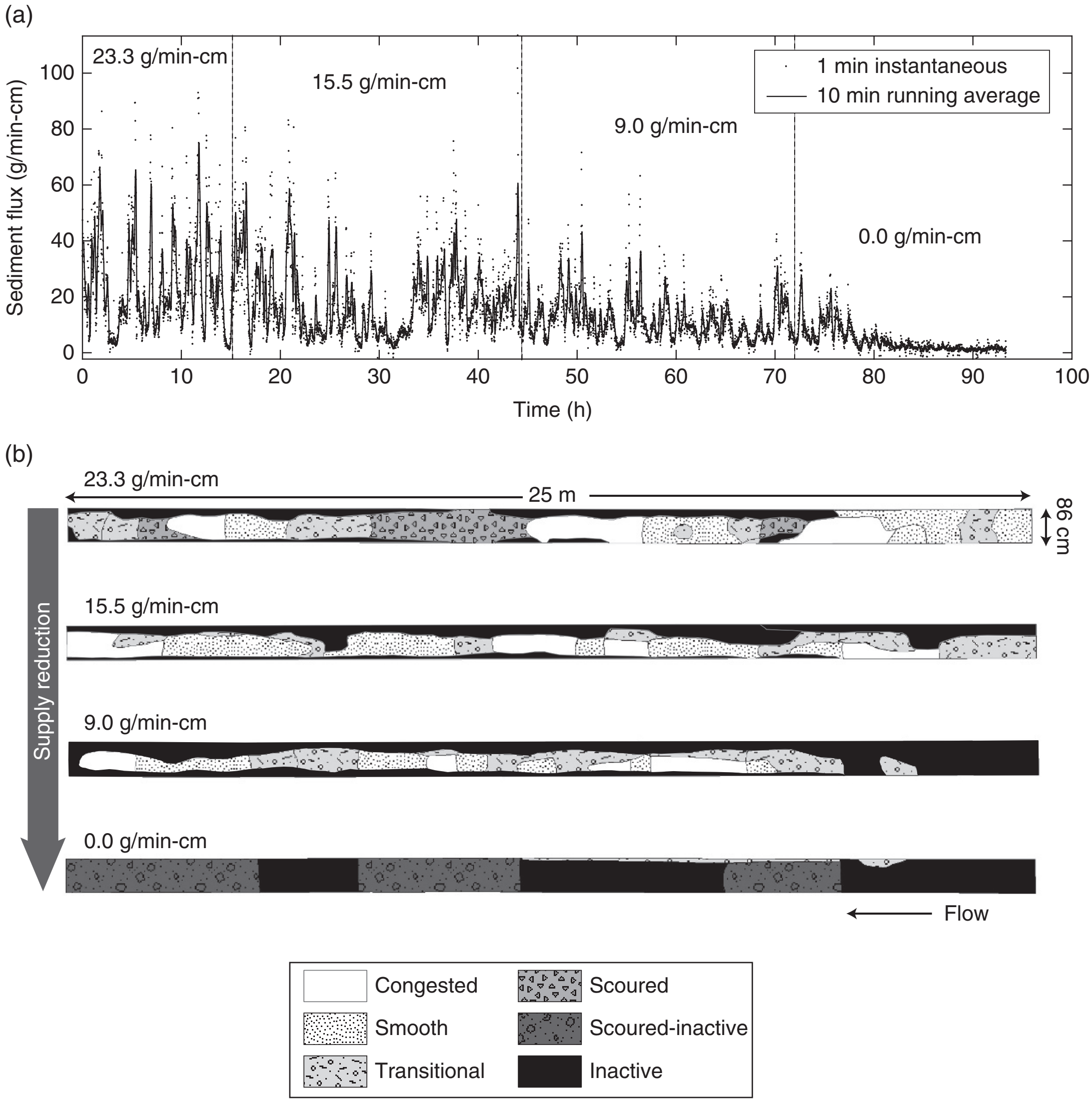


Figure 16.1 Response of (a) bedload transport and (b) bed-surface texture to reductions in sediment supply from 23.3 to 15.5 to 9.0 to 0.0 g/min-cm. At each supply rate, distinct patches were identified and classified as "congested" (coarse, D_{50} = 4.68 mm), "transitional" (D_{50} = 3.63 mm), "smooth" (fine, D_{50} = 2.62 mm), and "inactive" (coarse zones with no active bedload transport, D_{50} = 5.49 mm) following Iseya and Ikeda (1987). Bed areas that appeared to have a bimodal grain-size distribution were classified as "scoured," (because the intermediate-sized material was absent, exposing fine grains. See Nelson *et al.* (2009) for further details on the experimental procedure. (*Source*: Nelson *et al.* (2009). Reproduced with the permission of AGU.)

channel width and there is a small fraction of the bed surface grain-size distribution that can be mobilized only during extreme flood flows with higher transport capacity. Reduction of the sediment supply narrows the zone of active transport, the bedload fines and the mean bed surface grain-size becomes coarser (Dietrich *et al.* 1989; Lisle, Iseya, and Ikeda 1993; Nelson *et al.* 2009).

Dietrich *et al.* (1989) argued that the effects of sediment supply on the bed surface grain-size distribution can be quantified with respect to the transport capacity. Invoking a conventional functional form where the bedload transport rate is proportional to the excess boundary shear stress raised to a power (e.g., Meyer-Peter and Muller, 1948),

$$q_b = k(\tau - \tau_c)^{1.5}, \tag{16.1}$$

the transport rate over the coarse surface layer normalized by the transport rate over a surface as fine as the subsurface or load is

$$q_* = \left(\frac{\tau - \tau_{cs}}{\tau - \tau_{cl}}\right)^{1.5} \tag{16.2}$$

where q_b is the bedload transport rate per unit width, k is a constant, τ is the boundary shear stress, τ_c is the critical shear stress for particle entrainment and the subscripts s and l indicate values of τ_c for the bed surface and the load. Writing the shear stresses in nondimensional form

$$\tau^* = \frac{\tau}{(\rho_s - \rho)gD_{50}} \tag{16.3}$$

allows Equation 16.2 to be rewritten as

$$q_* = \left(\frac{\tau/\tau_{cl} - D_{50s}/D_{50l}}{\tau/\tau_{cl} - 1}\right)^{1.5} \tag{16.4}$$

where D_{50} is the median of the grain-size distribution. Dietrich *et al.* (1989) showed that when the sediment supply matches the river's ability to transport the load, q_* is unity. As the sediment supply decreases with respect to the transport capacity, the bed surface becomes increasingly armored, as in Figure 16.1. Figure 16.2 shows the variation in q_* with the ratio between the median size of the bed surface (D_{50s}) and load (D_{50l}) using a variety of data sets where a systematic, stepwise reduction of sediment supply was undertaken at a constant discharge. We assume that D_{50s}/D_{50l} is equivalent to the armor ratio under the assumption that the bedload and subsurface size distributions are the same. The data conform with the original observation from Dietrich *et al.* (1989) in that there is an asymptotic decline in q_* as the bed gets coarser and armors. Included in Figure 16.2 are data from Church, Hassan, and Wolcott (1998), whose experiments did not have a sediment feed, but instead started with a gravel–sand mixture and progressively armored at a constant flow. This data set does not show a clear asymptotic decline in q_* with the assumed armor ratio equivalent, but the authors do note that there was an asymptotic decline in sediment flux from the beginning of their experiment as the upstream bed armored and fed winnowed sediment to the downstream end of the channel where flux was measured.

Building on the underlying concept from Dietrich *et al.* (1989), Church and Hassan (2005) proposed that the value of q_* could be more easily estimated without assuming a specific functional form of the sediment transport equation by calculating $q_{*(C\&H)}$ as the ratio of the bedload transport rate for a bed surface that was armored after 96 hours of flow to the transport rate of the same bed in the first hour of their experiment, before it armored, at the same flow strength. It is assumed that the unarmored bed transport rate is approximately equal to the capacity for the grain-size distribution and hydraulic conditions (depth, slope, velocity). Extrapolating that concept, $q_{*(C\&H)}$ can be calculated for an experiment with sediment feed reductions as the bedload transport rate after the bed has adjusted to a new sediment feed divided by the transport rate for an unarmored bed, before any feed reductions.

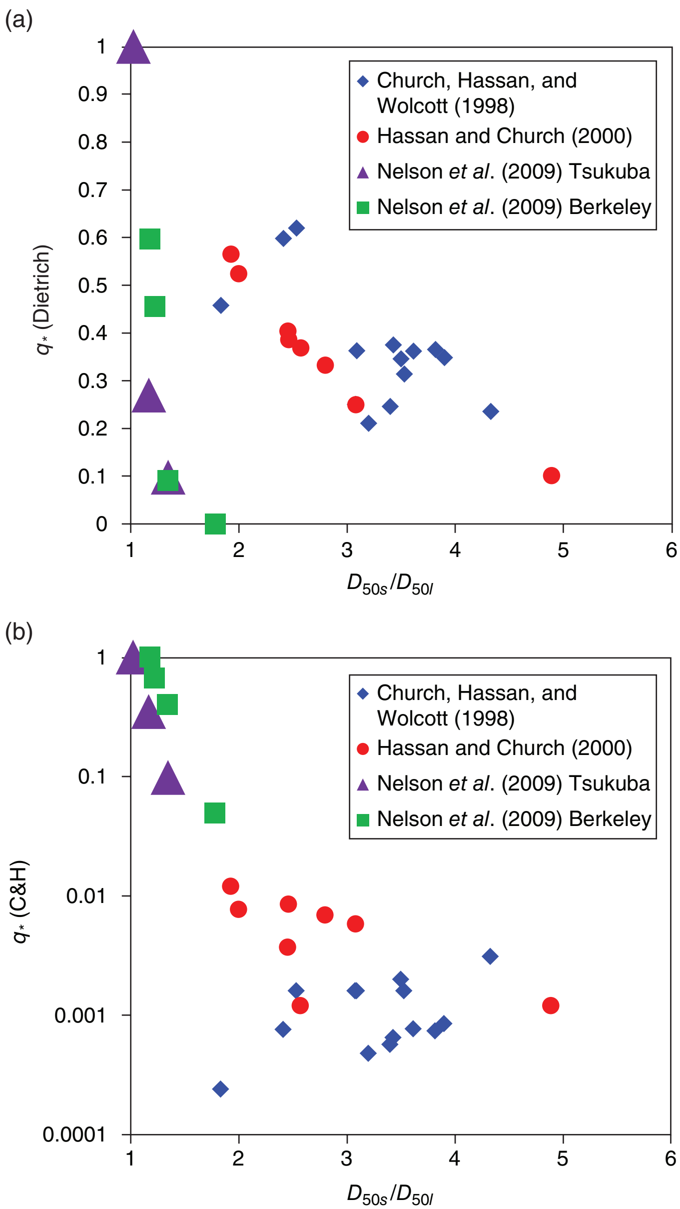


Figure 16.2 Variation in q_*, calculated using the (a) Dietrich *et al.* (1989) and (b) Church and Hassan (2005) methods, with the assumed armor ratio equivalent, D_{50s}/D_{50l}. Both experiments reported by Nelson *et al.* (2009) had bedload sheets with the exception of the lowest q_* value for the Berkeley experiments. Tsukuba data are the same data presented in Dietrich *et al.* (1989). The Church, Hassan, and Wolcott (1998) and Hassan and Church (2000) experiments both produced clusters, ribs, and reticulate cells.

Figure 16.2b shows the variation in $q_{*\,(C\,\&\,H)}$ with the assumed armor ratio equivalent, revealing the same asymptotic behavior as q_* with D_{50s}/D_{50l}. However, the q_* values are 2 orders of magnitude larger than the $q_{*\,(C\,\&\,H)}$ values for the Church *et al.* (1998) and Hassan and Church (2000) data. The difference between the two methods lies in the assumed functional form of the bedload transport equation in Dietrich *et al.* (1989). Church and Hassan (2005) show that transport rates in their experiments are overpredicted by ~2 orders of magnitude by the Meyer-Peter Muller Equation using a value of $\tau_c^* = 0.045$, so $q_* >> q_{*\,(C\,\&\,H)}$. They also suggest that the development of bed structures increases τ_c^* by ~50%. Examination of Figure 16.2 suggests that the experiments reported in Nelson *et al.* (2009), Church *et al.* (1998) and Hassan and Church (2000) all appear to form a continuous asymptotic decline in $q_{*\,(C\,\&\,H)}$ with increasing D_{50s}/D_{50l}. The Nelson *et al.* (2009) data reveal a sharp decline in $q_{*\,(C\,\&\,H)}$ and the Church *et al.* (1998) and Hassan and Church (2000) experiments form the stable asymptote. As discussed below, differences in the relative mobility of the grain size distributions in the Nelson *et al.* (2009), Church *et al.* (1998), and Hassan and Church (2000) experiments are probably responsible for the different behavior.

16.2.2 Mobility of the Grain-Size Distribution

The mobility of grain-size distributions affects the size of the material available for transport and therefore the bed's response to changes in sediment supply. Gravel mixtures can experience full, partial, selective, and equal mobility. There is some confusion in the literature over the precise definition of these terms, and universally accepted definitions have not been adopted. Here we follow the definitions of Parker (2008), which appear to capture all the possible behaviors. Partial transport occurs when the coarse tail of the bedload size distribution is finer than that of the bed surface (Figure 16.3a). Selective transport occurs when all grain sizes on the bed are found in the bedload, but the bedload size distribution is finer than the bed surface (Figure 16.3b). Equal mobility occurs when the size distributions of the bedload and the bed surface are the same (Figure 16.3c). Equal mobility of the bedload with respect to the subsurface sediment (rather than the surface) can occur in the presence of a coarse surface layer (e.g. Parker and Klingeman, 1982), which is a special case of selective mobility (Figure 16.3d).

Most gravel-bed streams exist in a condition of partial mobility during flows below bankfull, but have selective mobility during bankfull flows. True equal mobility, where the size distribution in transport is the same as the surface, occurs when the fraction of particle size i in the bedload p_i and the fraction of that size on the bed f_i are the same for all grain sizes. Essentially, for equal mobility the coarse surface layer must disappear. This condition is rare in gravel-bed rivers, a conjecture supported by the widespread use of hiding functions in sediment transport formulae to account for the fact that $p_i/f_i = 1$ is rare. Equal mobility with respect to the subsurface is more typically observed at high flows (Powell, Reid, and Laronne 2001) and can produce mobile armor layers (Wilcock and DeTemple 2005).

The true definitions of equal, selective, and partial mobility depend on the shape of the p_i/f_i curve with grain size (Figure 16.4). For equal mobility, the curve is flat across a range of grain sizes. For selective mobility, p_i/f_i declines for the large grain sizes, but p_i/f_i is greater than 0 for the largest grain size class, while for partial mobility, p_i/f_i is zero for larger grain sizes. It is not always practical to calculate p_i/f_i to determine the transport regime as this requires detailed measurements of both the bed surface and bedload grain-size distributions. Differential mobility conditions can be indexed by examining the ratio of the boundary shear stress to the shear stress required to move a particular grain size, τ/τ_{ci} (Wilcock and McArdell 1993). A practical choice for that grain size is the 84th percentile of the bed material (D_{84}) because it is diagnostic of whether the coarse tail of the grain-size distribution

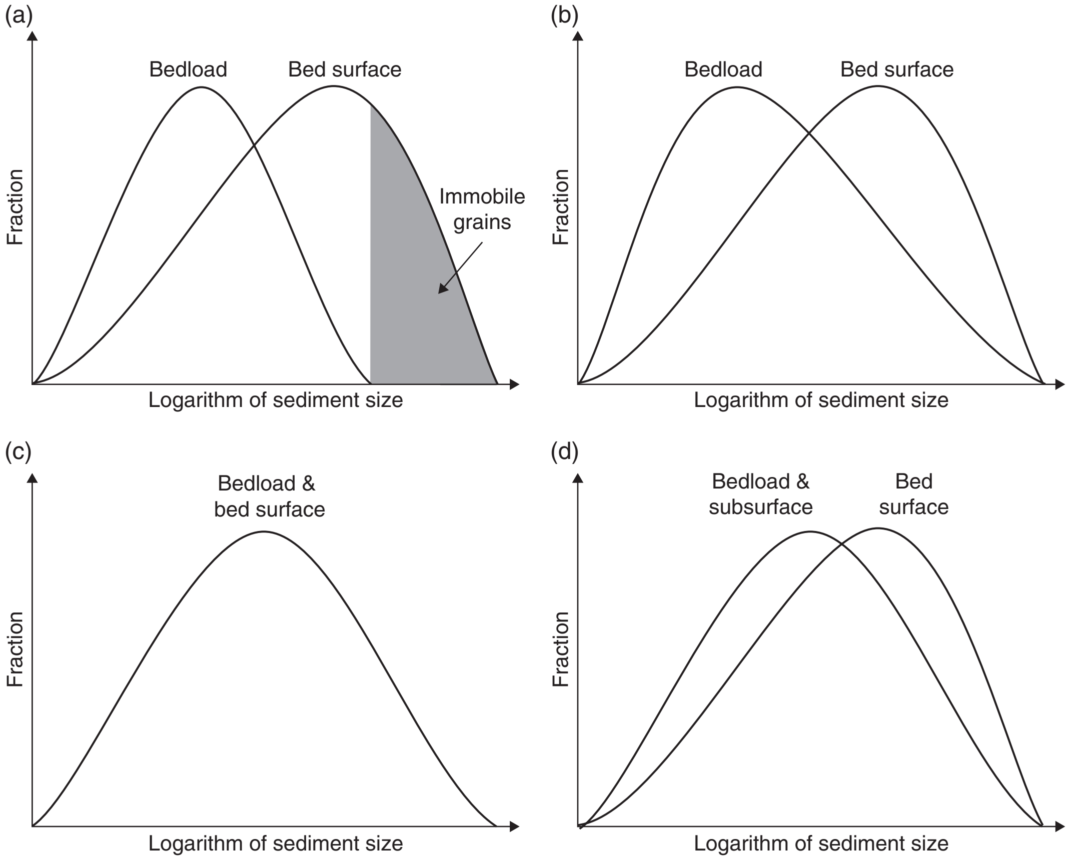
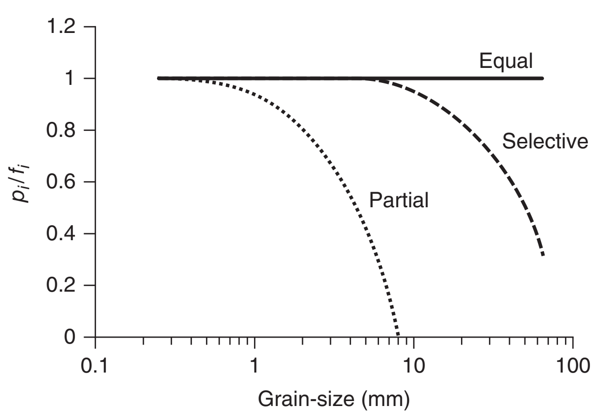


Figure 16.3 Hypothetical grain-size distributions for (a) partial, (b) selective and (c) equal mobility in gravel-bed rivers based on Parker (2008). Panel (d) depicts a special case of selective mobility where transport is equally mobile with respect to the subsurface sediment.

Figure 16.4 Conceptual illustration of how the ratio of fraction of grain size i in the bedload p_i and on the bed f_i varies with grain size for partial, selective, and equal mobility.



is moving. Partial mobility occurs when τ/τ_{c84} <1 because the shear stress is not competent to move the large grain sizes on the bed. For the same reason, selective mobility occurs when $\tau/\tau_{c84} > 1$. The arguments presented in Wilcock and McArdell (1993) suggest that when $\tau/\tau_{c84} > 2.1$, all grains are equally mobile. Some field studies have noted markedly higher values, which are thought to be linked to structuring of the bed surface material (cf. Powell *et al.* 2001).

While τ/τ_{c84} may be a useful criterion, it requires accurate characterization of shear stress applied to sediment grains on the bed, which is not easy because detailed measurements of the flow are needed as well as assumptions or measurements of the critical shear stress required to entrain a particular grain size. Furthermore, the differences reported for the critical values of τ/τ_{c84} for equal mobility make it impractical to use. An alternative is to cast τ/τ_{c84} into a ratio of the D_{84} for the load (or subsurface) and the bed surface. When $D_{84s}/D_{84l} = 1$, the grain size distribution is equally mobile. When the D_{84s}/D_{84l} >1, the grain-size distribution is selectively or partially mobile because large particles are underrepresented or absent in the load. To our knowledge, there is no theoretical D_{84s}/D_{84l} threshold at which conditions should change from selective to partial mobility. However, experimental conditions that exhibit strict partial mobility ($p_i/f_i = 0$ for the large grains), have values of D_{84s}/D_{84l} >2 (cf. Wilcock and McArdell 1993; Church, Hassan, and Wolcott 1998; Hassan and Church, 2000; Kleinhans *et al.* 2002), so we adopt this empirical threshold here to distinguish between selective and partial mobility.

16.3 Bed Features in Gravel-Bed Rivers

The types of bed features that emerge on a gravel bed are dictated by the sediment supply, relative to transport capacity, and the mobility regime. Figure 16.5 shows how the various types of bed features observed in gravel-bedded rivers plot in the phase space created by $q_{*(C\,\&\,H)}$ and D_{84s}/D_{84l}. Below, we

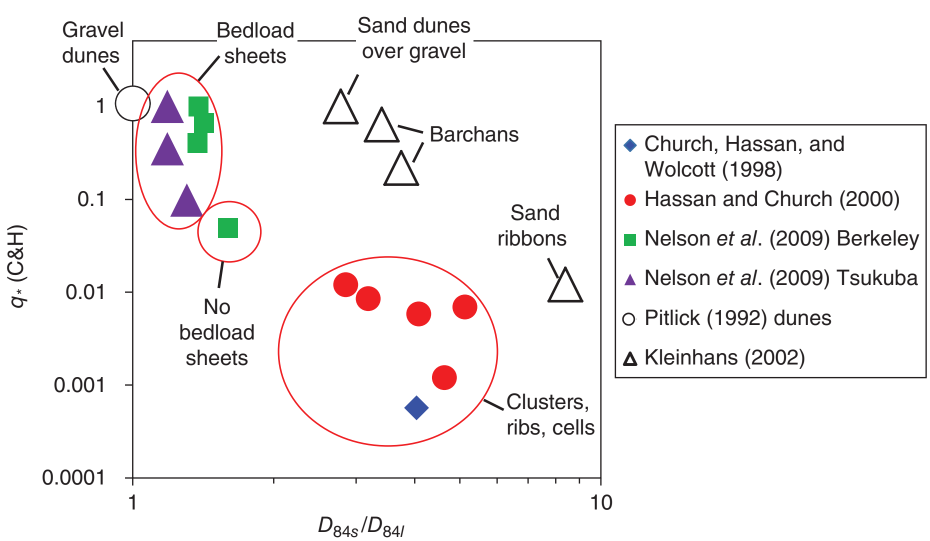


Figure 16.5 Bedforms developed under differing sediment supply and mobility regimes. Equal mobility occurs when $D_{84s}/D_{84l} = 1$, selective mobility conditions occur when $1 < D_{84s}/D_{84l} < 2$, and partial mobility conditions occur when $D_{84s}/D_{84l} > 2$.

provide some notes on the morphology and occurrence of each type of bed feature and attempt to elucidate the sediment supply and mobility conditions required for each bed feature.

16.3.1 Equal Mobility Regime

Figure 16.5 suggests that the bed features developed under equal mobility conditions in gravel-bedded rivers – where the supply equals the capacity – are gravel dunes. Here, we distinguish between gravel dunes and sand dunes developed in mixed size sediment where sand dunes may form over an otherwise immobile gravel bed (e.g. Kleinhans *et al.* 2002; Tuijnder, Ribberink, and Hulscher 2009) or sand dunes with a subsidiary gravel component (Blom, Ribberink, and De Vriend 2003), which we treat separately below. Early work on gravel transport argued that it was a flat-bed phenomenon and classical bedforms developed in sand-bedded channels were absent in gravel-sized sediments (see Carling (1999) and references therein). This idea is captured in many bedform phase diagrams showing that the dune field terminates at grain sizes larger than ~4 mm (e.g. Allen 1984). Yet at high shear stresses, gravel dunes have been observed to emerge from an otherwise flat bed during flood flows in the field (c.f. Dinehart 1989, 1992a,b; Pitlick 1992) and gravel dunes are commonly observed to develop over bars following flood flows (Figure 16.6; see also Baker 1984). Carling (1999) convincingly argued that this has led to misidentification of many features that are probably gravel dunes in both laboratory and field flows.

Gravel dunes have been identified in flows where the nondimensional Shields stress (τ_*; Equation 16.3) exceeds 0.1 (e.g. Hubbell *et al.* 1987; Kuhnle and Southard 1988; Dinehart 1989, 1992a,b; Carling 1999), which is approximately twice the typical entrainment threshold ($\tau_c^* = 0.045$) for gravel mixtures (Miller,

Figure 16.6 Gravel dunes formed on a bar in the Columbia River near Revelstoke, British Columbia. (Photograph courtesy of Rolf Kellerhals.)

McCave, and Komar1977; Yalin and Karahan 1979). Dinehart (1989, 1992a) described the growth of gravel dunes from incipient conditions that reach a maximum height at $\tau_* = 0.25$ and washed out to a flat bed at $\tau_* = 0.3$. This behavior is nearly identical to hydraulically-controlled sand dunes.

Carling, Richardson, and Ikeda (2005) showed that shear stress greatly in excess of the critical value for entrainment is not an absolutely necessary condition for fine gravel dune development, provided the flow persists for a long enough time. Similar observations have been made for dunes in sand-bedded rivers. Venditti, Church, and Bennett (2006) argued that bedforms are initiated instantaneously when the shear stress greatly exceeds the critical shear stress for entrainment, where the sand layer can be fluidized forming hydrodynamic instabilities between the flow and the bedload transport layer that are imprinted on the bed. They also noted that sand dunes can be initiated from defects in the bed when the flow is at or below the threshold of motion for the whole sediment bed, but the development process took much longer. Carling, Richardson, and Ikeda (2005) show a classic case of defect-type initiation of gravel dunes at shear stresses below where the bed can fluidize. However, gravel dunes would certainly be more common when formed during flood flows where the shear stresses are high enough for the transport layer to behave like a continuous fluid media so that they could form instantaneously.

Supply and Mobility Conditions

The sediment supply and mobility conditions for gravel dunes have never been described in the literature. The information provided in the classic references is insufficient to properly assess these parameters primarily because of the difficulty in measuring bedload and the bed-surface grain size simultaneously. However, data reported by Pitlick (1992) for the North Toutle River provides some insight into the conditions necessary to produce gravel dunes. We used Pitlick's (1992) reported bedload transport measurements and shear stress to calculate the bedload capacity of an unarmored bed from the Meyer-Peter and Muller (1948) equation to determine a value of $q_{*\,(C\,\&\,H)}$. We also used bed material D_{84}, measured ~6 months after the bedload measurements to estimate the ratio D_{84s}/D_{84l}, under the assumption that this value of D_{84} was characteristic of the channel-bed surface. The calculations show that, on average for the dune beds, $q_{*\,(C\,\&\,H)} = 1.08$ and $D_{84s}/D_{84l} = 0.99$ (Figure 16.5). We attempted the same calculations using a number of other data sets (notably Dinehart 1992a) and found the only reported bed-surface measurement (Dinehart 1992b) was much finer than the bedload, suggesting that unlike the Pitlick (1992) measurement, it was not characteristic of the channel at higher flows. This suggests that gravel dunes form under conditions where the supply equals or nearly equals the capacity and the transport condition is near equal mobility. Furthermore, gravel dunes develop under the same conditions as sandy bedforms, which are hydraulically controlled insofar as their emergence and dimensions can be predicted from the mean hydraulics in a channel (i.e., depth, slope, velocity, and shear stress).

16.3.2 Selective Mobility Regime

The bed features that plot in the selective mobility range of Figure 16.5 ($1 \le D_{84s}/D_{84l} \le 2$) are bedload sheets. First identified by Whiting *et al.* (1988), they appear in river channels as bands of the coarser fractions of the bed material that grade upstream to the finer fractions. These features are commonly aligned across the primary flow path in river channels and migrate downstream, but it is not uncommon to see bedload sheets shoaling onto bars as a result of secondary flow patterns (Wooldridge and Hickin, 2005). The formative mechanism for bedload sheets has not been fully elucidated, but the sorting patterns across bedload sheets suggest that they pose abrupt changes in roughness and turbulence structure (Best 1996; and references therein) and Seminara, Colombini, and Parker (1996) proposed that the stress perturbation due to this sorting structure allows for the growth of bedload sheets.

Bedload sheets are widely thought to be a precursor to gravel dune development (Bennett and Bridge 1995a,b) when flows are sufficiently large that grain inertia is overcome to begin the vertical stacking process required for dune growth. This can only happen quickly in flows well in excess of the critical threshold (Carling 1999). Some authors regard bedload sheets as analogous to dunes (cf. Bridge 1993), but the evidence for this conjecture is not yet well supported. The numerous attempts to include bedload sheets into the bedform phase diagrams used to predict bedform occurrence in sandy sediments (cf. Best 1996; Carling 1999; Kleinhans *et al.* 2002) have only been partially successful at distinguishing between bedform types in poorly sorted sediments. Indeed, bedload sheets commonly fall into the dune existence field on bedform phase diagrams. This has provided some evidence of the linkage between bedload sheets and dunes formed in coarse sediments (gravel and sand–gravel mixes).

Bedload sheets migrate downstream as a consequence of the "catch and mobilize" process, in which large grains are caught in the wakes of other large grains, followed by infilling of their interstices by smaller particles, which can in turn smooth out hydraulic wakes causing large particles to be remobilized (Whiting *et al.* 1988). The low-amplitude coarse front of a bedload sheet develops when a sufficient concentration of coarse particles accumulates (so-called "gravel jams"; Iseya and Ikeda 1987). Relative to the finer sheet tail, this coarse front is hydraulically rough and has relatively large friction angles (e.g., Buffington, Dietrich, and Kirchner 1992; Kirchner *et al.* 1990), which provide distrainment sites where other particles are likely to become temporarily deposited. As fine particles fill the interstices of this coarse front, the bed becomes smoother and near-bed flow accelerates, which disproportionately increases the drag on the coarse particles because they are exposed further into the flow than fine particles (Venditti *et al.* 2010). This causes the coarse particles to become remobilized, and once they are plucked from the bed, they tend to roll over finer bed material until they once again become caught in a coarse sheet front.

The remobilization of coarse particles through grain interactions with fine particles and consequent migration of bedload sheets in gravel–sand mixtures agrees conceptually with observations that the presence of sand can increase the mobility of gravel (e.g., Iseya and Ikeda 1987; Ferguson, Prestegaard, and Ashworth 1989; Wilcock 1998; Wilcock, Kenworthy, and Crowe *et al.* 2001; Wilcock and Kenworthy 2002; Wilcock and Crowe 2003; Curran and Wilcock, 2005). Indeed, most field and flume observations of bedload sheets have occurred under conditions with gravel–sand mixtures (e.g., Iseya and Ikeda 1987; Whiting *et al.* 1988; Kuhnle and Southard 1988; Dietrich *et al.* 1989; Wilcock 1992; Bennett and Bridge 1995a; Bunte *et al.* 2004; Kuhnle *et al.* 2006; Madej *et al.* 2009). However, Nelson *et al.* (2009) showed that bedload sheets, and the catch-and-mobilize process, can occur in a unimodal gravel without any sand present. Thus it seems likely that it is the ratio of coarse to fine particles in the distribution, rather than the presence of sand, that determines whether bedload sheets are able to develop and migrate.

The dynamics (wavelength and celerity) of bedload sheets are controlled in large part by the rate of sediment supply. Nelson *et al.* (2009) showed that under conditions of high sediment supply, bedload sheets tended to move downstream more quickly and sheets were spaced more closely together than those observed under low supply conditions. They suggested that these phenomena can be explained by simple mass balance arguments where, because the fronts of bedload sheets tend not to grow beyond a height of 1–2 coarse grain diameters, the bedform celerity is proportional to the sediment transport rate (i.e., the sediment supply), and the contributing bed area necessary to provide enough coarse particles to form the sheet front (that is, the space between consecutive bedload sheets) shrinks with increasing sediment supply.

Supply and Mobility Conditions

The migration process of bedload sheets precludes the possibility that they are formed under a partial mobility condition because the whole grain-size distribution must experience transport in order to pass the waveform downstream. This suggests that channels with bedload sheets should exhibit an equal mobility condition, integrated over the waveform, with respect to the sediment supply, along the corridor of active bedload transport. Because the zone of active transport narrows with decreasing sediment supply as immobile coarse patches expand on the channel margins (Lisle, Iseya, and Ikeda 1993; Nelson *et al.* 2009), under low to moderate sediment supply conditions the grain-size distribution of the whole bed surface will likely be coarser than the distribution of the load and the channel will therefore be in a state of selective mobility. If the sediment supply becomes so low that the coarse patches extend across the entire width of the channel, bedload sheets should stop moving and gradually disappear.

We see this behavior in the Nelson *et al.* (2009) experiments with bedload sheets formed in pure gravel (Figure 16.1). As sediment supply to a channel is reduced, the zone of active transport narrows, bedload sheet migration rate declines until, under conditions of zero sediment supply, the sheets eventually disappear and the bed becomes uniformly coarse (Nelson *et al.* 2009). Had sand been part of the bed material mixture, it is likely the bed at this point would have entered a partial mobility condition as the sand was winnowed from the bed. But because the bed was composed entirely of gravel, the bed surface grains became interlocked against one another, trapping the fine material below the surface (Wydgza *et al.* 2005). This highlights a degree of freedom that generally does not exist in a sand-bedded channel. If the supply to a sand-bedded channel is reduced, the bed will erode to compensate for the sediment flux divergence. In a gravel-bedded channel, erosion of the bed will also occur, but as the bed erodes, the bed can become coarser under selective mobility conditions reducing and stabilizing the erosion.

The forgoing argument and Figure 16.5 appear to suggest that bedload sheets form under selective mobility conditions when the ratio of sediment supply to transport capacity for the subsurface is $0.1 < q_* < 1$. What happens if the transport regime is equally mobile rather than selective? It would be expected that incipient dunes begin to form (e.g. sediment feed experiments reported in Kleinhans 2002).

16.3.3 Partial Mobility Regime

There are two distinct types of bedforms that plot in the partial mobility regime of Figure 16.5. Partial-mobility conditions can occur with relatively high sediment supply ($q_* > 0.1$) when sand is transported over an immobile armored gravel bed. Under low sediment supply ($q_* < 0.1$) conditions in gravel mixtures, bed structures develop.

16.3.3.1 High Sand Supply Bedforms (q_* >0.1)

At high values of q_*, supply-limited sandy bedforms can develop over an otherwise immobile gravel bed. This is a special condition of partial transport whereby the gravel-bed armor is developed, but finer material continues to be supplied to the channel bed. This can occur through a gravel–sand transition where sand is supplied to the bed from suspension (e.g. Venditti and Church 2014; Venditti *et al.* 2015), when coarse gravel supply to the channel is episodic but sand supply from the floodplain and hillslopes is persistent, or where there has been a permanent shift in sediment supply from gravel to sand, and thus the gravel bed is in the process of being buried. The bedforms that develop under conditions of a static coarse bed and a fine sediment supply are well documented and include sand-ribbons, barchans, and dunes with gaps in the troughs that progressively close as the sediment supply increases until there is a fully alluvial sand bed (Figure 16.7a; Kleinhans *et al.* 2002; Tuijnder, Ribberink, and Hulscher 2009; Tuijnder and Ribberink 2012). Figure 16.7b–e shows the results of a simple phenomenological experiment where sand was fed to a 1-m wide Plexiglas channel, revealing that as the sand supplied to an otherwise sediment-starved flow increases, the sequence of bedforms arises even under the simplest experimental conditions, without additional roughness elements. This sequence of bedforms is not limited to mixed gravel–sand-bedded rivers, as it has been well-documented in eolian environments (Bagnold 1941; Lancaster 1995), tidal environments (Allen 1968; Ernstsen *et al.* 2005) and on the sea floor (Lonsdale and Malfait 1974; Lonsdale and Spiess 1977), suggesting that a definable suite of bedforms develops under sand supply-limited conditions in many flow environments at the Earth's surface.

There has now been extensive work in flumes on this sequence of bedforms over otherwise immobile gravel beds in laboratories (e.g. Kleinhans *et al.* 2002; Grams 2006; Tuijnder, Ribberink, and Hulscher 2009; Tuijnder and Ribberink 2012) and the field (Kleinhans *et al.* 2002; Venditti *et al.* 2009). There has also been some work on dune development in mixed size sediments where there is a subsidiary gravel component in the bed mixture that forms an armor layer in the dune troughs, but also plays a role in active transport over the dunes (e.g. Blom, Ribberink, and De Vriend 2003). Generally, the work has shown that sand supply to the bed controls the bedform morphology (Figure 16.7a) and, in particular, it is the thickness of the sand coverage that sets the types of bedforms that emerge (Kleinhans *et al.* 2002; Venditti *et al.* 2009). Bedforms also grow with transport stage in addition to sediment supply (Figure 16.7a). Kleinhans *et al.* (2002) recognized this and cast their results in the form of a phase diagram defined by a bed mobility parameter (transport stage) and a ratio of the predicted to measured bedload, which markedly improved the distinction between bed features where sand was being transported over immobile gravel.

Supply and Mobility Conditions

It is clear from the existing literature that this sequence of sandy bedforms developed over an otherwise immobile bed is characteristic of the partial mobility regime because the gravel bed is generally immobile and sandy bedforms are migrating over it. Unfortunately, there have not been any systematic experiments exploring these features where the sediment supply and transport rate are independent, as occurs in sediment feed flumes. Most work on these types of bedforms in mixed gravel–sand mixtures has been undertaken in sediment recirculating flumes (e.g., Kleinhans *et al.* 2002; Blom, Ribberink, and De Vriend 2003; Kuhnle *et al.* 2006) where the sediment supply is controlled by the transport rate in the flume, allowing a partial mobility condition to persist in equilibrium (Wilcock and McArdell 1993). If a gravel–sand mixture is fed into a flume, and the flow is only competent to carry the sand, the gravel will cause aggradation until the slope of the channel can pass the incoming supply of all grain sizes, eventually shifting the transport condition to a selective mobility condition where bedload sheets are more likely to form, at least along a narrow active transport corridor (Figure 16.1). So whether this distinctive suite of bedforms can form with a mixed sand–gravel sediment supply is unclear. Data presented in Kleinhans (2002) from a sediment feed flume did not reveal the sequence of bedforms observed in his recirculating flume experiments; instead, he found sand ribbons and bedload sheets coexisted in the selective mobility regime and mixed sand–gravel dunes developed as the transport conditions approached equal mobility. Hence, we speculate that sandy bedforms developed over an otherwise immobile gravel bed can be sustained only if the gravel supply is terminated.

Within this context, our placement of the Kleinhans *et al.* (2002) data in Figure 16.5 should be viewed with some caution. In order to increase the sand supply, they increased the shear stress to disturb the armor layer and release more sand from the bed until sand dunes developed. In our calculation of $q_{*\,(C\,\&\,H)}$, we regarded the fully sand-bed transport rate as the unarmored transport rate and

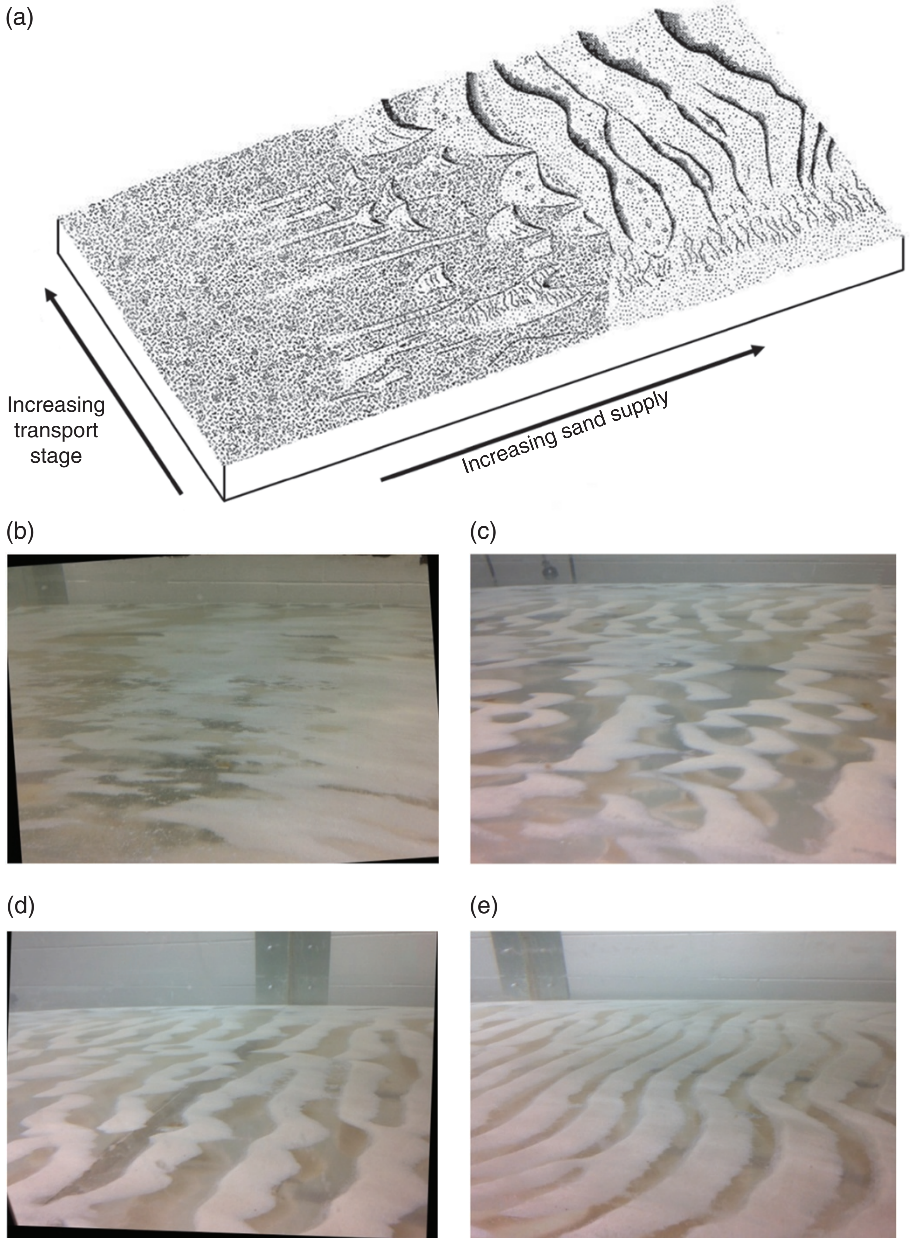


Figure 16.7 (a) Kleinhans *et al.*'s (2002) conceptual model of sand bedforms developed over an otherwise immobile gravel bed showing the influence of transport stage and sediment supply: flow is left to right. (Modified from Kleinhans *et al.* 2002). Sequence of bedforms developed of over a Plexiglas bed with increasing sand supply showing (b) patches of sand akin to sand ribbons, (c) barchan dunes, (d) interconnected barchan dunes, and (e) laterally continuous dunes: flow is right to left.

subsequent states (with lower shear stresses) as the transport rate after the bed had armored. Nevertheless, the data conform to the other patterns observed in Figure 16.5. At low values of $q_{*\,(C\,\&H)}$, sand ribbons form. Similar patches of sand have been described as beds are winnowed of finer particles (e.g. Hassan and Church 2000). As $q_{*\,(C\,\&\,H)}$ increases, the barchans form, and then when $q_{*\,(C\,\&\,H)} \approx 1$, sand dunes form with an underlying gravel armor layer. The bed material is still composed of a gravel–sand mixture; however, the experiment remained in the partial mobility regime. If transport stage increases in a channel, and the armor layer can be entrained, a shift to selective mobility dunes could occur (e.g. Blom, Ribberink, and De Vriend 2003), but these would be hydraulically controlled features.

16.3.3.2 Low Gravel Supply Bed Structures (q_* <0.1)

The bed features that plot in the partial mobility range when $q_{*\,(C\,\&\,H)}$ <0.1 are structural features of armor layers that include "pebble" clusters (Dal Cin 1968; Brayshaw 1984,1985; Strom and Papanicolaou 2009), transverse ribs (McDonald and Banerjee 1971; McDonald and Day 1978; Koster 1978) or stone lines (Laronne and Carson 1976), and reticulate stone cells described by Church, Hassan, and Wolcott (1998) and others (e.g. Gustavson 1974; Laronne and Carson 1976; Hassan and Church 2000). A clear taxonomy for these structural features of a gravel bed has not been presented in the literature, so for simplicity, we adopt the terms clusters, ribs, and reticulate cells.

Cluster bedforms are closely nested groups of clasts aligned roughly parallel with the flow and are widely considered the dominant microform roughness in gravel-bedded rivers (Dal Cin 1968; Brayshaw 1984). They are thought to form on the falling limb of hydrographs when the largest stones in the bed material, keystones, making up the D_{95} of the grain size distribution, stop moving. These keystones lie above the mean bed elevation of the surface armor and accumulate particles larger than the D_{75} on the upstream stoss side and particles smaller than the D_{50} in the lee of the keystone (Brayshaw 1984) forming the characteristic morphology shown in Figure 16.8a, d, and e. Brayshaw *et al.* (1983) shows that the lee deposit forms as particles are captured in the keystone wake and interacting flow fields can lead to particle capture on the stoss side. However, it has also been shown that clusters can be formed by particle interactions, without consideration of the fluid dynamics (Tribe and Church 1999). The features tend to be longer in the streamwise direction than in the cross-stream direction, are not regularly spaced, and have been observed to occupy 5–10% of the total bed area. Cluster size varies with the caliber of the sediment, but clusters up to 1.2 m in length have been documented (Brayshaw 1984). The stoss-side deposit forms as shear stress declines below the threshold for movement of the larger stones, and the lee-side deposit develops when the conditions permit deposition of the D_{50}. The flow field developed around a keystone is critical to the interaction of the large stones in the bed material because once a moving clast is caught in the upstream perturbed flow field of the keystone or its wake, it is essentially removed from the sediment load (Brayshaw, Frostick, and Reid 1983; Malmaeus and Hassan 2002).

The bed material comprising a cluster can typically be reentrained only by movement of the keystone. Billi (1988) describes a process whereby as flows increase, the lee particles are entrained, followed by the stoss particles, followed by dislodgement of the keystones at high flow. As such, cluster bedforms wash out at high flows and are reestablished on the declining limb of hydrographs (Billi 1988; De Jong 1991; Strom and Papanicolaou 2009), although continued disaggregation has been documented on the falling limb of hydrographs (Reid, Brayshaw, and Frostick 1984). Aggregation and disaggregation of groups of particle clasts is a fundamental component of gravel transport processes. Strom *et al.* (2007) show that clusters can form in uniform sediments, and it is common to see temporary clusters form as part of the transport process in mixed sized sediments. These temporary clusters formed during the transport process are distinguished from the more persistent cluster bedforms created under partial mobility conditions where the largest particles on the bed are immobile.

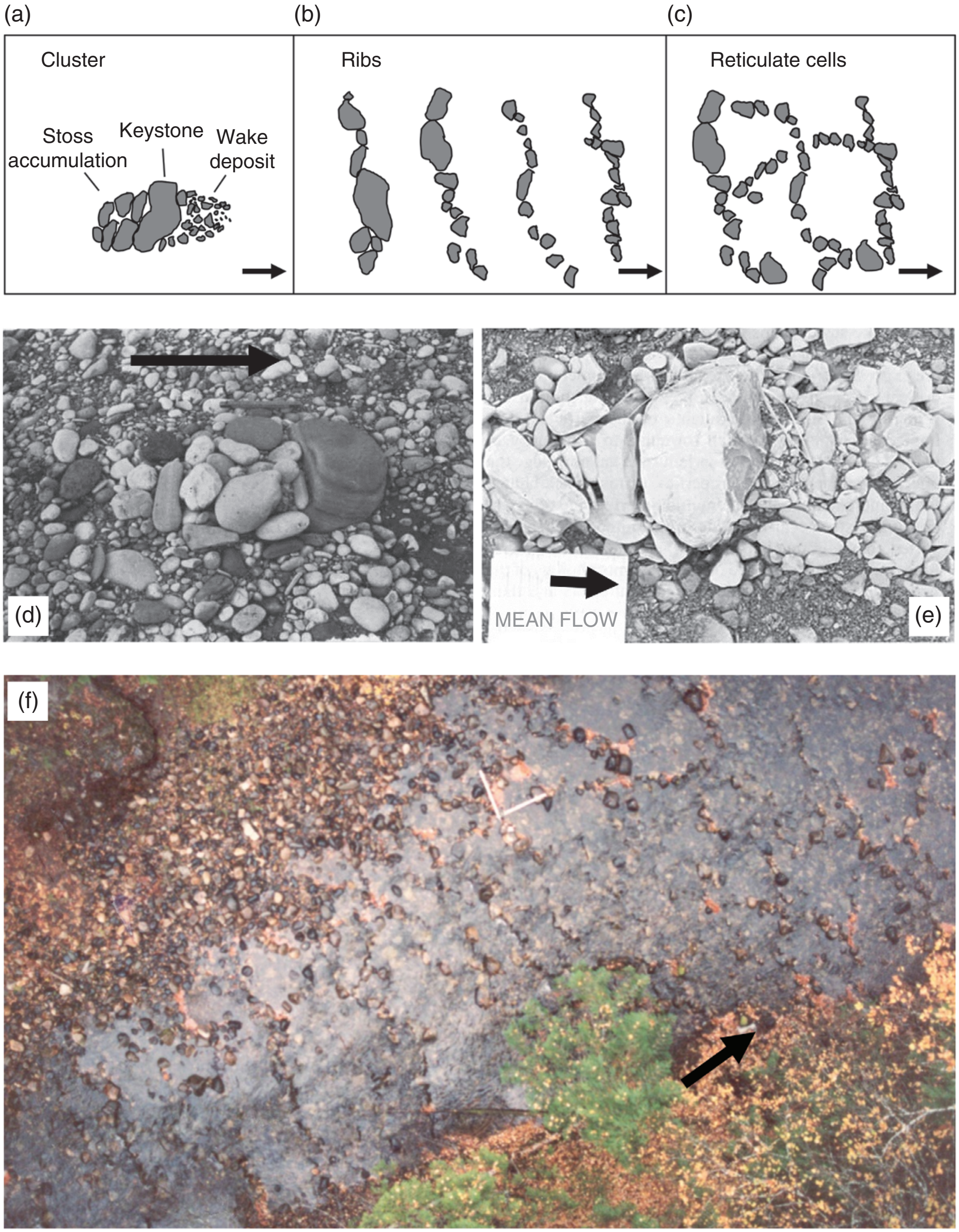


Figure 16.8 (a–c) Planview cartoons of the structural features of an armored surface, including (a) clusters, (b) ribs, and (c) reticulate cells. In panels b and c, only the large grains on the bed (> D_{84}) are depicted. Planview photographs of (d) a gravel cluster from Piave River, Italy, (e) a gravel cluster from Afon Elan, Wales, and (f) ribs and reticulate cells on the bed of Harris Creek, British Columbia, Canada. Flow direction is highlighted in all panels by the black arrows. ((d) *Source*: Cin 1968. Reproduced with permission of Elsevier; (e) from Brayshaw 1984, and (f) photograph courtesy of Michael Church.)

After clusters develop, intermediate features known as stone lines or transverse ribs may form (Figure 16.8b and f). There are considerable differences in the sedimentology and genetic relation among the various features that have been described as transverse ribs. Morphologically, they have been described as regularly spaced rows of pebbles, cobbles, or boulders lying transversely across the primary flow direction (McDonald and Banerjee 1971; Gustavson 1974; McDonald and Day 1978; Koster 1978; Allen 1983, 1984; Church, Hassan, and Wolcott 1998). Like clusters, transverse ribs are typically not more than a few grain diameters high and scale with the clast sizes. They typically extend across the flow for lengths greater than their alongstream spacing. The formative process for transverse ribs has been linked to supercritical flow, where the ribs develop under stationary hydraulic jumps (Koster 1978; McDonald and Day 1978; Allen 1983, 1984). This explanation may be plausible, but it requires hydraulic jumps to be stationary in space and time long enough for large particles to organize, which seems unlikely in a natural channel. Furthermore, some transverse ribs described in the literature look like the reticulate cells described by Church, Hassan, and Wolcott (1998), others may be bedload sheets, and still others are obviously step-pool morphologies (see examples provided by Macdonald and Banerjee (1971) and McDonald and Day (1978)), none of which are linked to supercritical flows. Step-pool morphologies appear to be linked to the jamming of and interlocking of large stones, on steep slopes where the flow plays a subsidiary role in setting the length and hydraulic roughness of the step-pool morphology (Zimmerman and Church 2001; Church and Zimmerman 2007), but their stability is argued to be related to the stochastic arrangement of clasts in the channel (Zimmerman, Church, and Hassan 2010). Our use of the term ribs is to describe the class of transverse features that form in gravel-bedded rivers under supply-limited conditions and excludes more exotic features such as transverse lines of pebbles formed over a silt bed (Macdonald and Banerjee 1971) or transverse lines of vegetation formed on concrete (Allen 1984), and channel-scale, step-pool morphology (Macdonald and Banerjee 1971).

Church, Hassan, and Wolcott (1998) described cluster bedforms as an intermediate stage in the development of stable reticulate cells, which form as a mixed-size bed develops an armor layer (Figure 16.8c). They consider these structures a stable self-organized product of the armoring process whereby particles move from less stable positions to more stable positions. Church, Hassan, and Wolcott (1998) argue that reticulate cells are derived from ribs formed under subcritical flow conditions. The cluster growth between the ribs interconnects them to form a web of cellular structures. Laronne and Carson (1976) describe similar features as closed structures that can trap finer-grained material. The features described by Church, Hassan, and Wolcott (1998) are on average 10 D_{84} long and 6 D_{84} wide. Structural development varies with the sediment transport, which is reasonable because higher transport rates require more frequent exchanges between mobile sediment and the bed. If larger particles are moving, the clasts forming the reticulate structures will be entrained. Hassan and Church (2000) reported experiments where they fed sediment to an otherwise immobile bed and found that greater sediment supply rates resulted in more weakly developed cells.

Some insight into the controlling mechanisms of bed structures has been acquired by simplified numerical modeling. Tribe and Church (1999) simulated the development of clusters, ribs, and reticulate cells using a simple rules-based approach and showed that these features may be formed by grain-to-grain interactions, without any consideration of the flow. In their model, particles stop on contact with other grains and may cause those grains to move. The density of large stones in the simulation domain controlled the development of different structures. When the domain had a coverage of 5–10% of large stones, linear along-stream cluster-like features developed. When the coverage was 20–30% of the domain, a loosely transverse series of interconnected ridges developed that were similar to reticulate cells, and when the coverage was > 40%, rib features emerged. When grains were allowed to rotate, transverse lines became more tightly packed if an entrainment probability rule (smaller grains are more likely to be entrained) and a neighbor rule (stone is entrained only if it has few close or touching neighbors) were implemented. Tribe and Church (1999) concluded that grain rotation and domain coverage were the primary controls on the emergence of reticulate cells observed by Church, Hassan, and Wolcott (1998), with the entrainment probability and the neighbor rule reinforcing the tight packing. A more probabilistic approach that includes entrainment thresholds based on shear stresses, resistance fields about particles for distrainment, a vertical protrusion-dependent entrainment rule, and a particle dropping rule to prevent vertical accretion, has produced more realistic simulations (Malmaeus and Hassan 2002).

Supply and Mobility Conditions

Clusters are formed by particles that have stopped moving, thereby allowing the accumulation of other clasts on the stoss side and the formation of a fine wake deposit. This would suggest they are a feature associated with partial mobility conditions. Similarly, ribs and reticulate cells are built from these clustered particles during the armoring process, so they too are characteristic of the partial mobility regime. In fact, the development of bed structures enhances partial mobility conditions by making it more difficult to entrain coarse particles. Field and flume measurements of interlocked coarse grains have shown that the vertical plucking force required to remove a grain from a static, dry bed can be 3–5 times the weight of the grain (Wyzdga *et al.* 2005). Feeding sediment to the bed at even nominal amounts leads to less well-defined cellular structures (Hassan and Church 2000), indicating they should be formed when sediment supply is low relative to the capacity.

16.4 A Phase Diagram for Bed Features in a Gravel-Bedded River

In sand-bedded rivers, the emergence of various bedforms have been effectively linked to measures of flow strength (e.g. velocity, bed shear stress, or nondimensional shear stress) and grain size (see Southard and Boguchwal 1990; and reviews in Allen (1984), Southard (1991) and Venditti (2013)). Bedform phase diagrams based on the mean flow velocity and grain size are particularly good discriminators of bedform types in sand-bedded rivers where there is no sediment supply limitation. Other combinations of flow strength and grain size are less effective. For example, bedform phase diagrams based on the shear stress and grain size often exhibit overlap of several bedform types in the phase space (Venditti 2013). There have been attempts to place bed features of gravel-bedded rivers on bedform phase diagrams developed for sand-bedded rivers with varying success. Carling (1999) added gravel dunes and bedload sheets to the mean velocity–grain size phase diagrams from Southard and Boguchwal (1990) and bed shear stress–grain size phase diagram of Allen (1984). The exercise is important in the present context, because it shows gravel dunes are hydraulically controlled. Indeed, if depth is taken into account in the Southard and Boguchwal (1990) diagram, gravel dunes plot in the same phase space (extended for larger grain sizes) as sand dunes. However, there is concerning overlap between plane-bed, dunes, and bedload sheets across the phase space on both diagrams, which prevents their use as a predictive tool in gravel-bedded rivers.

Placing the bedforms, structures, and patches that develop in gravel-bedded rivers on the flow-strength-based diagrams developed for sand-bedded rivers explicitly assumes flow strength exerts first-order control on these bed features. While we cannot discount the effect of flow strength, the discussion above and Figure 16.5 suggest that a more useful phase diagram may be formed by variables representing the sediment supply (relative to the transport capacity) and the differential mobility of

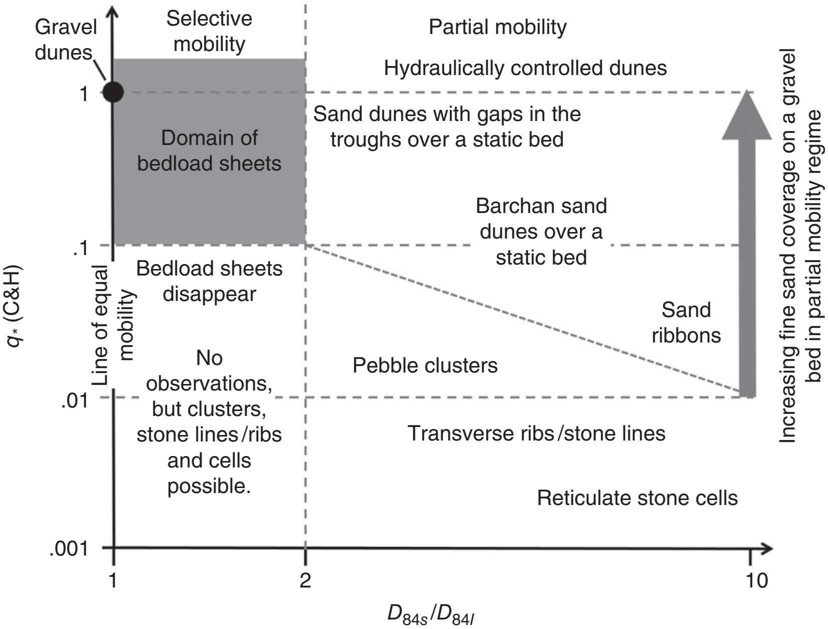


Figure 16.9 Conceptual bedform phase diagram for gravel beds, partly based on Figure 16.5.

particles in a mixture. Both of these are dependent on flow strength, but we assert that the flow strength itself plays a secondary role in the types of bed features that emerge.

Figure 16.9 shows a conceptual phase diagram for bed features developed under sediment supply-limited conditions in gravel-bedded rivers. The placement of the various features is based partly on Figure 16.5, but is also constrained by the above discussion of the sediment supply and mobility conditions required for the formation and stability of each bed feature. The phase diagram shows our proposed line of equal mobility at $D_{84s}/D_{84l} = 1$, selective mobility conditions between $1 < D_{84s}/D_{84l} <2$ and partial mobility condition when $D_{84s}/D_{84l} >2$. The bed is aggradational when $q_* >1$. The tradeoff between surface armoring and erosion in response to the sediment flux divergence created when $q_* <1$ is difficult to assess qualitatively, but a slow reduction in sediment supply, as is typical on the waning phase of hydrographs, would favor surface coarsening rather than bed incision and this is borne out by experimental work (Dietrich *et al.* 1989; Church, Hassan, and Wolcott 1998; Hassan and Church 2000; Nelson *et al.* 2009). It is of course possible to have $D_{84s}/D_{84l} <1$, but this should only occur under aggradational conditions where the bed surface is being buried by coarser particles, so conditions to the left of $D_{84s}/D_{84l} = 1$ and $q_* <1$ should not be physically possible.

These conditions define unique existence fields for bedforms, mobile patches, and structures in gravel-bedded rivers. We show gravel dunes occurring under equal mobility conditions when the supply is closely matched by the transport capacity. The dune area we identify on the phase diagram is a single point, which is too restrictive. For example, dunes observed in Kleinhans' (2002) sediment feed experiments would plot in the selective mobility region near the line of equal mobility, but uncertainties regarding the nature of the bed surface prevent us from including them in Figure 16.5. Furthermore, there is no reason why gravel dunes cannot form under aggradational conditions. In fact, their presence in the rock record (Carling 1999) dictates this. Nevertheless, our fundamental argument here is that gravel dunes are hydraulic phenomena and not developed by sorting processes, and our placement of them on this diagram is commensurate with that idea. Figure 16.9 is therefore best suited to highlighting bed features developed under supply-limited conditions.

The domain of bedload sheets lies within the selective mobility regime and we have indicated that they occur when $q_* > 0.1$, based on the experimental data presented in Nelson *et al.* (2009). They show that as sediment supply is reduced and the bed coarsens, the patchiness responsible for bedload sheet migration disappears. What happens to bedload sheets when $q_* << 0.1$ is not clear as there are no experimental data in that range. The bed may simply become static, but it is more likely that after bedload sheets disappear, the bed will shift to a partial mobility regime where the coarsest stones on the surface become immobile and the sequence of clusters, ribs, and reticulate cells emerge. The transport conditions described by Church, Hassan, and Wolcott (1998) as producing these bed features persisted during the final, zero-feed phase of the Nelson *et al.* (2009) Berkeley experiments when the coarse particles became interlocked and generally immobile (Wydzga *et al.* 2005). A logical question is whether bedload sheets can occur under aggradational conditions. Bennett and Bridge (1995a,b) and Madej *et al.* (2009) both report archetypical bedload sheets under aggradational conditions. This extends the domain of bedload sheets beyond our nominal placement based on Figure 16.5 to values of q_* up to ~1.5.

The clusters, ribs, and reticulate cells described by Church, Hassan, and Wolcott (1998) and Hassan and Church (2000) dominate when $q_* << 0.1$. Our placement of these bedforms in a sequence from higher to lower q_* and offset diagonally is purposeful. Hassan and Church (2000) describe the reticulate structures as being best developed when there is essentially no sediment supply and the bed has fully armored. Therefore, it is reasonable to expect that clusters will form first as the bed armors. As the bed continues to armor and the transport becomes increasingly partially mobile, transverse ribs will form. Finally, once the sediment transport rate has reached a stable asymptote, the development process is complete and the reticulate cells emerge as the dominant feature of the bed structure. Nevertheless, the available data do not distinguish between these features and we have taken some artistic liberty in their placement on the phase diagram.

At higher values of q_* and when D_{84s}/D_{84l} is large, Kleinhans *et al.* (2002) report sand ribbons emerging as the dominant bedform on the coarse surface layer. In a partial mobility regime, an increase in q_* represents increasing fine sand coverage on a gravel bed, which leads to the formation of barchans. Sand dunes with gaps in the troughs over a static bed form as q_* is further increased until the bed becomes aggradational with fine sediment and transitions to a fully alluviated sand bed with hydraulically controlled bedforms. Our placement of the sequence of sand ribbons, barchans, and dunes with gaps in the troughs diagonally is purposeful. Our expectation is that as q_* approaches 1, the grain-size distribution will become less partially mobile (more of the larger grains will move) and may transition to a stable condition of hydraulically controlled, mixed-size sand dunes with coarse troughs and finer stoss slopes and crests (Blom, Ribberink, and De Vriend 2003).

16.5 Perspective and Conclusions

We have reviewed the impact of sediment supply on bedload transport rates and the emergence of various bedforms, structures, and patches in gravel-bedded channels. Sediment supply, relative to capacity, appears to play a critical role in the emergence and disappearance of bed features and a distinct suite of features develops under equal, selective, and partial mobility conditions. Dunes appear to be the dominant bedform under equal mobility conditions where the sediment supply is closely matched by the transport capacity and they can form under aggradational conditions. Bedload sheets occur under conditions of selective transport when the grain-size distribution is wide enough to allow coarse particles to be mobilized by the fine fraction. Under the special case of a fine sediment supply to an otherwise coarse immobile gravel bed (partial mobility), a sequence of sandy bedforms (sand ribbons, barchans, and dunes with gaps in the troughs) will emerge until the fine sediment supply overwhelms the bed and it becomes sand-bedded. When the sediment supply to a gravel bed is low compared to the capacity in the partial mobility regime, the bed evolves a sequence of clusters, transverse ribs, and reticulate cells as the bed coarsens. These represent the end product of the coarsening process and the organization of the armor layer in a gravel-bedded channel.

Our framework provides an opportunity to explore controls on local bed features regardless of the source of the sediment supply. In the laboratory experiments we examined, the sediment feed provided the sediment supply. In the field, sediment supply is typically derived from the upstream reach, where it may be controlled by hydraulics. However, sediment supply may also be controlled by upstream or within-reach hillslope processes (landslides, debris flows, direct particle delivery) or bank erosion. Our framework does not distinguish between these sources because local bed features are responding to whatever is provided from immediately upstream.

Our framework also simplifies a complex process in gravel-bedded channels: variable discharge. Because our framework focuses on grain-size distributions and relative sediment supply, it is insensitive to variable flow. The bed features record the last flow capable of moving particles. As such, we could use the dominant bed features and grain-size distributions of the bed surface and subsurface (assumed to be equivalent to the load) to estimate the relative sediment supply conditions in a channel without knowing the flow. This could then be inverted to estimate the flow if supply could be estimated from downstream sediment deposits that capture all of the bedload. Our framework could therefore be developed into a paleoflow estimator. However, some consideration needs to be given to determine whether the predicted flow is the flow of interest. As with sandy bedforms, it is not always clear what aspects of variable flow (peak magnitude, duration of flow, steepness of rising and falling limbs of the flow) are responsible for the development of the bed features we observe in the field.

There are some impediments to applying the framework we developed. There are timescales of bed adjustment to consider when sediment supply is changed or changing, especially when the upstream sediment supply is controlled hydraulically. For sandy bedforms over a gravel bed, the timescale of adjustment is dependent on their size. For gravel clusters, the timescale of adjustment can be very short; clustering is an inherent process of gravel transport, so a sharp recessional hydrograph would leave clusters on a bed. Better developed stoss and lee deposits and interlocking caused by particle vibration would result from longer recessional hydrographs with long periods of partial mobility. Ribs and reticulate cells also become better developed through time as particles jostle into position on the recessional limb of hydrographs.

The greatest impediment to using the framework we developed here is the absence of data to verify the phase diagram. Our attempt to form a phase diagram from observations has been hampered by the availability of data to populate a diagram. Many experimental arrangements have used sediment-recirculating flumes (or partial-recirculating flumes where sand is recirculated and gravel is not) where the sediment supply is intricately intertwined with the channel hydraulics. Essentially, the sediment supply is always equivalent to the transport capacity, so q_* cannot be calculated reliably. While much can be learned about the transport process from these sediment recirculating experiments, these data are difficult to capture within our framework. Further hampering our ability to populate a phase diagram is that in order to quantify equal, selective, and partial mobility conditions, we need to know the behavior of the coarsest fractions of the bedload with respect to their presence on the bed. Few existing publications report sufficient information to do so. This is in part because many of the early publications on phenomena in gravel-bedded rivers documented features not previously reported and pre-date the pioneering work of Wilcock and collaborators on the differential mobility of grain-size distributions. However, we examined over 75 publications and found just four with the necessary information, a few with enough information for us to estimate parameters, and another few with enough information to intuit where they might lie on a phase diagram. This points to the need for a community data repository to facilitate storage and dissemination of existing and new sediment experimental data in a common format, which has recently been discussed extensively at workshops and scientific meetings (e.g., Hsu *et al.* 2015).

A better understanding of sediment supply controls on bed features in gravel-bedded rivers and verification of our phase diagram can ultimately be achieved with new field observations and laboratory experiments specifically designed to simultaneously measure bedload flux, the grain-size distributions of the bed surface and bedload, and bed morphology under different sediment supply rates for a variety of gravel and gravel–sand mixtures. We expect such work will result in refinement of the ideas presented herein and bring our understanding of bed features in gravel-bedded rivers to the same level that we understand bedforms developed in sand-bedded rivers.

Acknowledgments

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Discussion

Discussion by Jie Qin

The author presents an impressive phase diagram which represents a promising framework to investigate bed features emerged on gravel surfaces. This diagram is an improved version of another diagram presented by the author (Venditti, Nelson, and Dietrich 2008). In both diagrams, the authors argue that the emergence of the bed features is controlled by sediment conditions (e.g. sediment supply and grain-size distribution), while hydraulic conditions play a subsidiary role in their development. I agree that sediment conditions play an important role in developing the bed structures as demonstrated in previous literature, but the relative importance of sediment conditions and hydraulic conditions may need to be elaborated. Supposing that there is a simple recirculating flume experiment filled with heterogeneous gravel materials, the flow discharge increased gradually from zero until reaching an equal mobility situation. During this experiment, different bed features can be readily observed, while, in my opinion, these features are mainly determined by hydraulic conditions or by both hydraulic and sediment conditions.

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Discussion by Jonathan Laronne

The attempt to construct phase diagrams for bedforms smaller than the channel scale but larger than individual grains excepting dunes in gravel-bedded rivers is very interesting. The main argument is that "the emergence of these features is controlled by sediment supply and the differential mobility of grain-size distributions (equal, selective, and partial mobility)." If supply is the controlling mechanism, ideally a given river reach or large bedform such as a bar would be characterized by one bedform. Nonetheless, except for starved gravel-bedded rivers, all the rest are characterized by a wealth of bedforms of the type examined in these flume runs. They form intricate patches of more mobile and less mobile bed materials. As such, sediment supply must be very locally controlled and not reach-controlled. This also attests to the fact that facies of fluvial conglomerates include a variety of fabrics and grain size distributions within a given unit.

Response by the Authors

Most gravel-bed rivers exhibit discrete patches of similar grain size that may be free (migrating sorting features), fixed (spatially persistent sorting features caused by weak, grain-scale, topographic controls and local coarsening) or forced (spatially persistent features due to strong topographic controls such as bar morphology and channel obstructions) (Nelson *et al.*, 2009, 2010). The occurrence of these patches is related to reach-scale sediment supply, insofar as channels without a supply will eventually become uniformly coarse and immobile. Our phase diagram considers the features developed due to local sediment supply conditions. It does not consider the variability in bed features that would emerge across channel-scale topographic features such as bars, steps, and pools due to spatial gradients in local sediment supply. Nevertheless, it is likely that there are characteristic sequences or patterns of bed features that are related to these channel-scale topographic features, which directly reflect reach-scale sediment supply conditions. Identification of these reach-scale patterns remains a problem for further research.

The relative role of hydraulics and sediment supply in creating bed features is not fully resolved. In uniform sediment, bedforms that emerge are controlled by the interaction of a turbulent flow field with the labile bed and sediment supply is not an important consideration so long as it is high enough to keep the bed fully alluviated. In non-uniform sediment, the bed surface grain-size distribution adjusts to sediment supply (Dietrich *et al*. 1989). As such, the quantity and caliber of sediment supplied from upstream exerts a dominant role in what types of bed features emerge in response to a particular flow. If flow is held constant and upstream sediment supply is varied, the type of bed feature that emerges depends on the sediment supply (see Figure 16.1). The flow is only important insofar as it is responsible for the type of transport condition it generates (partial, selective, or equal) for a particular size distribution. Sediment recirculating flumes exhibit fundamentally different behavior than sediment feed flumes in that the quantity and caliber of sediment supplied is dependent on flow strength because the sediment that exists the flume is what is supplied to the upstream end (cf. Parker and Wilcock, 1993; Wilcock and DeTemple, 2005). In this case, the types of bed features that emerge may superficially appear to be controlled by hydraulic conditions, but ultimately hydraulics is setting the sediment supply, which in our framework, sets the bed feature.

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