Errata: Author corrections shown in blue for Table 32.1 and Fig. 32.6.

32

Changes in Channel Morphology Over Human Time Scales

John M. Buffington

32.1 INTRODUCTION

Rivers are exposed to changing environmental conditions over multiple spatial and temporal scales, with the imposed environmental conditions and response potential of the river modulated to varying degrees by human activity and our exploitation of natural resources. Watershed features that control river morphology include topography (valley slope and channel confinement), discharge (magnitude, frequency, and duration of runoff events), sediment supply (volume, calibre and frequency of sediment delivery), and vegetation (riparian communities (bank strength, roughness) and in-channel wood debris). River stability and response to changing environmental conditions are highly dependent on local context (channel type and associated degrees of freedom; the nature of the imposed sediment, hydrologic, and vegetation regimes; imposed anthropogenic constraints; and the legacy of past natural and anthropogenic disturbances).

Understanding the processes responsible for channel change and assessing river stability has long been of interest to geologists and engineers (Gilbert, 1877; du Boys, 1879; Davis, 1889; Lindley, 1919; Shields, 1936; Lane, 1937; Mackin, 1948; Leopold and Maddock, 1953; Schumm, 1969). Over the last several decades, interest in this topic has grown exponentially as a result of environmental legislation that has spurred greater interdisciplinary collaboration amongst physical and biological scientists studying riverine ecosystems and watershed processes. During this time, considerable progress has been made in understanding channel response and in elucidating biophysical interactions.

For example, progress has been made in understanding textural and structural response of gravel streambeds and consequent effects on bed mobility (Dietrich *et al.*, 1989; Church *et al.*, 1998; Wilcock, 1998; Buffington and Montgomery, 1999a; Nelson *et al.*, 2009) that, in turn, affect the availability of riparian and aquatic habitats (Lisle, 2005; Burke *et al.*, 2006; May *et al.*, 2009; Moir *et al.*, 2009). Sediment routing models have been developed for examining a wide range of responses, including textural adjustment, reach-scale changes in sediment storage and channel slope, changes in channel width and planform, effects of sediment pulses, and landscape evolution (Hoey and Ferguson, 1994; Nicholas *et al.*, 1995; Benda and Dunne, 1997a, 1997b; Cui and Parker, 2005; Coulthard *et al.*, 2008; Parker *et al.*, 2011; Mosselman, Chapter 9, this volume). In addition, our knowledge of the form and function of rivers has extended beyond traditional studies of sand- and gravel-bed rivers to include a wide variety of channel types and geomorphic settings; in particular, a tremendous amount of work has been done on step-pool channels (Comiti and Mao, Chapter 26, this volume). Numerous recent investigations also document the role of vegetation in affecting channel morphology and response over multiple spatial and temporal scales, including controls on channel width, grain size, bedforms, roughness, sediment transport, alluviation, and rates of landscape lowering (see review by Montgomery *et al.*, 2003). Recent studies also document the role of animals in modulating geomorphic processes and channel characteristics (Rice *et al.*, Chapter 19, this volume). Furthermore, historical studies demonstrate the startling extent of human disturbance to rivers worldwide (Schumm, 1977; Collins *et al.*, 2003; Surian and Rinaldi, 2003; Nilsson *et al.*, 2005; Chin, 2006; Surian, 2006; Walter and Merritts, 2008), fueling the notion that an anthropogenic epoch has emerged (Crutzen, 2002; Meybeck, 2003).

This chapter reviews channel change over human time scales (10^{-1}–10^{2} a), which encompasses small-scale adjustments resulting from seasonal changes in water-

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shed inputs to large-scale changes in reach morphology resulting from infrequent floods or decadal to centennial changes in climate (wet/dry cycles). The chapter begins with a review of scales of channel change and the spatial and temporal variability of channel response. This is followed by an examination of the available approaches for predicting channel change which, despite recent advances in the field, are still mainly limited to lowland rivers and assumptions of equilibrium conditions. Transient channel responses are explicitly accounted for in numerical models, but temporal variability of channel condition is typically absent in other quantitative models of channel response. Quantifying temporal variability is critical for assessing channel condition, planning restoration design, and predicting effects of climate change on riverine ecosystems. It is hypothesized that hydroclimate, and its control on both the shape of the hydrograph and the relative size of floods, is a first-order control on the natural variability of channel morphology.

32.2 SCALES OF CHANNEL CHANGE

Rivers can exhibit a broad range of responses to changing inputs of water, sediment, and vegetation over human time scales. Channel response may range from small-scale adjustment of channel characteristics (grain size, width, depth) to large-scale alteration of reach morphology and planform pattern.

32.2.1 Changes in Channel Characteristics

Successive, overlapping, spatial and temporal scales of morphologic response in alluvial channels include:

(1) Grain-scale adjustment, comprising:
   (a) local changes in grain size, packing, protrusion, and friction angle (Fenton and Abbott, 1977; Church, 1978; Buffington *et al*., 1992; Johnston *et al*., 1998);
   (b) development of micro-grain forms (e.g., particle clusters, stone cells; Laronne and Carson, 1976; Brayshaw, 1984; Church *et al*., 1998; Strom and Papanicolaou, 2008);
   (c) formation of textural patches (i.e., grain-size facies; Dietrich *et al*., 1989; 2006; Rice, 1994; Paola and Seal, 1995; Buffington and Montgomery, 1999a; Laronne *et al*., 2001; Nelson *et al*., 2009; Yuill *et al*., 2010).

(2) Changes in the type, size, and frequency of bed topography, ranging from micro-bed forms (e.g., ripples, bedload sheets; Whiting *et al*., 1988) to macro-bed forms or channel units (individual bar, pool, step, and riffle topography; Jackson, 1975; Lewin, 1978; Bisson *et al*., 1982; Church and Jones, 1982; Wood-Smith and Buffington, 1996; Halwas and Church, 2002; Hassan *et al*., 2008).

(3) Altered channel geometry (changes in local cross-sectional width, depth, and downstream variation of those features; e.g., Trimble, 1997).

(4) Altered stream gradient due to reach-scale aggradation/incision and changes in channel sinuosity. Here, stream gradient is distinguished from valley gradient, which is not adjustable in the short term.

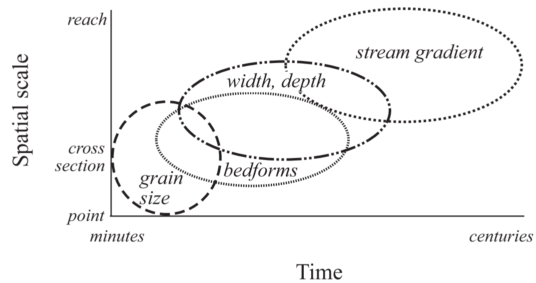

Figure 32.1 Spatial and temporal scales of channel response variables in alluvial rivers: grain size (individual grains to textural patches); width, depth (cross-section to subreach variation in average channel geometry); bedforms (micro-forms (e.g., ripples, bedload sheets) to channel units (e.g., bar, pool, step, riffle)); and stream gradient (reach-scale channel aggradation/incision or change in sinuosity). Partly after Knighton (1998).

Larger scales of channel response reflect the cumulative action of smaller-scale processes, particularly sediment transport of bed and bank materials. Hence, a progression of successive scales of response can be envisioned, with grain-size adjustment being the first-order response (Figure 32.1). Furthermore, because alluvial rivers exhibit mutually adjusting channel characteristics, changes in any one parameter can influence all of the others.

The extent of channel change that occurs for a given disturbance depends on the amount of work accomplished by the event (flood magnitude times duration; Wolman and Miller, 1960) and the time needed for a given scale of response to occur. Wolman and Miller's (1960) classic magnitude–frequency argument emphasizes the effectiveness of frequent, moderate-sized events in accomplishing geomorphic work over the long term, but large-scale changes in morphology require large events (Figure 32.2). Similarly, different temporal scales of disturbances (seasonal to centennial) will exhibit characteristic scales of response. Seasonal changes are frequent, typically small-magnitude events that will lead to similarly small degrees of channel change (e.g.,

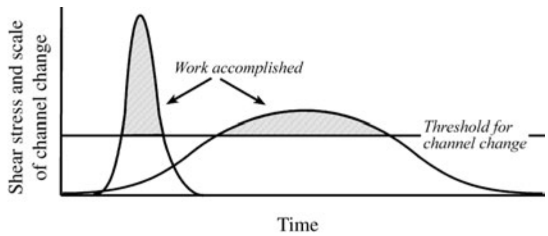

Figure 32.2 Work and scale of channel change accomplished for flood events of different magnitude and duration (after Costa and O'Connor, 1995). Total work (shaded area) is identical for the events, but the scale of channel change is greater for the high-magnitude event.

bed loosening and changes in grain-size structure and texture as seasonal floods begin; Milhous, 1973), while annual peak floods are typically moderate-sized events that lead to moderate scales of channel change (e.g., altered cross-sectional widths and depths). Over decadal scales, rare infrequent events (e.g., 50–100-year floods or debris flows) may cause significant channel change followed by a period of relaxation from the disturbance (recovery or attainment of some new equilibrium state; Bull, 1991; Simon and Rinaldi, 2006). The duration of the relaxation period depends on the magnitude of the disturbance and the local channel context (i.e., transport capacity relative to sediment supply; Montgomery and Buffington, 1997; 1998). For example, field observations show that steep, confined channels in central Idaho recover from massive debris-flow inputs within 5–10 years because of high transport capacities and no opportunities for floodplain storage of sediment, while recovery from sediment waves may take decades in lower-gradient, unconfined channels (Megahan *et al*., 1980; Madej and Ozaki, 1996). Over longer time scales (decades to centuries), wet/dry climate cycles may lead to corresponding cycles of river incision/aggradation (Rumsby and Macklin, 1994), expansion/contraction of channel width (Schumm and Lichty, 1963), and changes in the complexity of bed topography and textural patches. Overall, the sequence of different types and scales of disturbance will influence channel morphology and future responses (Hoey, 1992) with systems potentially showing complex and variable responses over time (Coulthard and Van De Wiel, 2007). Furthermore, the entire fluvial system may be trending toward some new state (i.e., non-stationarity) driven by longer-term changes in climate or by long-term anthropogenic alteration of the basin (Bryan, 1925; Schumm, 1968; 1977; Knox, 1984; Bull, 1991; Brierley and Fryirs, 2005).

The style and extent of channel response will also be limited by channel type and associated degrees of freedom (Table 32.1). In general, bedrock channels have fewer degrees of freedom for morphologic response than alluvial channels. In turn, steep, confined alluvial channels (e.g., step-pool and cascade streams) are less responsive than lower-gradient unconfined channels (e.g., pool-riffle and dune-ripple streams). Furthermore,

Table 32.1 Reach-scale channel type and degrees of freedom for morphologic change. (+ = likely; p = possible; – = unlikely)[a]

Channel type	Channel change				
	Grain size	Width & depth	Bedforms	Stream gradient (sinuosity/elevation)[b]	
Colluvial	p	p	–	–/p[c]	
Bedrock	–	–	–	–/–	
Cascade	p	–	–	–/–	confined
Step-pool	p	–/p[d]	p	-- /p	
Plane-bed	+	p/+	–	–/p	
Braided	+	+	+	+/+	
Pool-riffle	+	+	+	+/+	unconfined
Dune-ripple	p	+	+	+/+	

[a] Modified from Montgomery and Buffington (1997). Response potential for each channel type is discussed elsewhere (Montgomery and Buffington, 1997; 1998; Buffington *et al*., 2003).
[b] Changes in stream gradient may occur via altered sinuosity or incision/aggradation that alter the absolute elevation change across a reach. Slashes in the table distinguish these two responses.
[c] Fluvial incision/deposition is possible, depending on the degree of colluvial fill.
[d] Changes in channel depth can occur via pool fill/scour.

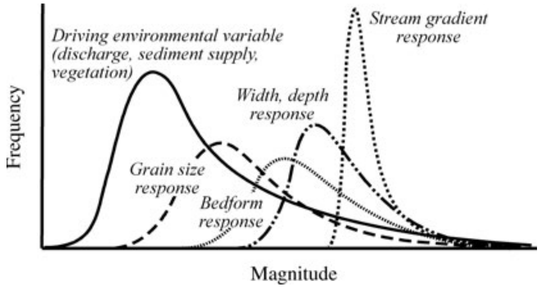

Figure 32.3 Channel change depends on the overlap between frequency distributions of driving and resisting forces for different scales of morphologic response.

because most geomorphic processes exhibit thresholds for occurrence, channel response will depend on the probability of a given disturbance exceeding the response threshold, the magnitude of which also varies with channel type. Hence, the likelihood of a given channel change will depend on the degree to which the probability distributions of driving environmental conditions (changes in discharge, sediment supply, vegetation) overlap with the probability distributions of process thresholds that lead to changes in channel characteristics (i.e., overlap of driving versus resisting forces; Figure 32.3).

32.2.2 Bed Mobility and Channel Type

Bed mobility is a first-order control on channel response and can be indexed by comparing the bankfull Shields (1936) stress (τ^*_{bf}) to the critical value for incipient motion of the median grain size (τ^*_{c50}). The bankfull Shields stress is defined as $\tau^*_{bf} = \rho ghS/[(\rho_s - \rho)gD_{50}]$, wherein ρ and ρ_s are fluid and sediment densities, respectively, g is gravitational acceleration, h is bankfull depth, S is channel slope, and D_{50} is median surface grain size. The critical Shields stress (τ^*_{c50}) for sand- and silt-bed rivers is determined here from the original Shields curve as fit by Brownlie (1981), while for coarse-grained rivers τ^*_{c50} is defined as a function of channel slope based on recent studies (Lamb *et al.*, 2008; Pitlick *et al.*, 2008); $\tau^*_{c50} = 0.15S^{0.25}$ and $0.36S^{0.46}$, respectively (also see Comiti and Mao, Chapter 26, this volume). Data compiled from rivers around the world show two features regarding τ^*_{bf} values (Figure 32.4). First, τ^*_{bf} varies with channel type, with a clear difference in behaviour for sand-bed channels versus coarser-grained rivers (also see Dade and Friend, 1998; García, 2000; Parker *et al.*, 2003; Church, 2006). The sand–gravel transition likely defines the boundary between the two (Figure 32.4; Sambrook Smith and Ferguson, 1995; Parker and Cui, 1998; Ferguson, 2003; Singer, 2008). Multithread channels are

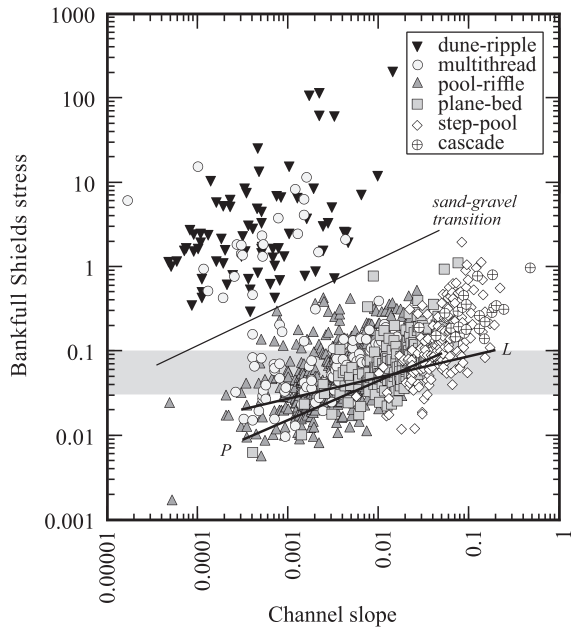

Figure 32.4 Bankfull Shields stress as a function of channel slope for different reach-scale channel types: dune-ripple (defined by morphology or assumed for single-thread channels with $D_{50} < 2$ mm; $n = 79$), multithread (divided/wandering, braided, and anastomosing channels, $n = 129$), pool-riffle (defined by morphology, or assumed for meandering channels with $D_{50} > 2$ mm; $n = 570$), plane-bed ($n = 185$), step-pool ($n = 168$), and cascade ($n = 23$). Where bankfull information was unavailable, other channel-forming flows were used (e.g., the two-year or mean annual flood). Data sources are reported in Section 32.9, Appendix. An approximate boundary for the sand–gravel transition is shown, as well as curves for critical Shields stress (τ^*_{c50}) reported by Lamb *et al.* (2008) and Pitlick *et al.* (2008); L and P, respectively, both limited to their respective range of observations. The shaded area indicates the range of τ^*_{c50} values reported by Shields (1936).

unique in that they span the sand- and gravel-bed domains of τ^*_{bf}. As defined here, this channel type encompasses divided/wandering rivers (Desloges and Church, 1989; Ferguson and Werritty, 1991), braided, and anastomosing channels; a lumped category is used because of inconsistent classification amongst the available data sources. The second feature of Figure 32.4 is that in coarser-grained rivers, τ^*_{bf} varies systematically with channel slope (Buffington and Montgomery, 2001; Mueller *et al.*, 2005; Pitlick *et al.*, 2008). However, the latter relation is potentially spurious, so it should not be taken as significant, at least on face value. Nevertheless, it is a convenient way to display the τ^*_{bf} values.

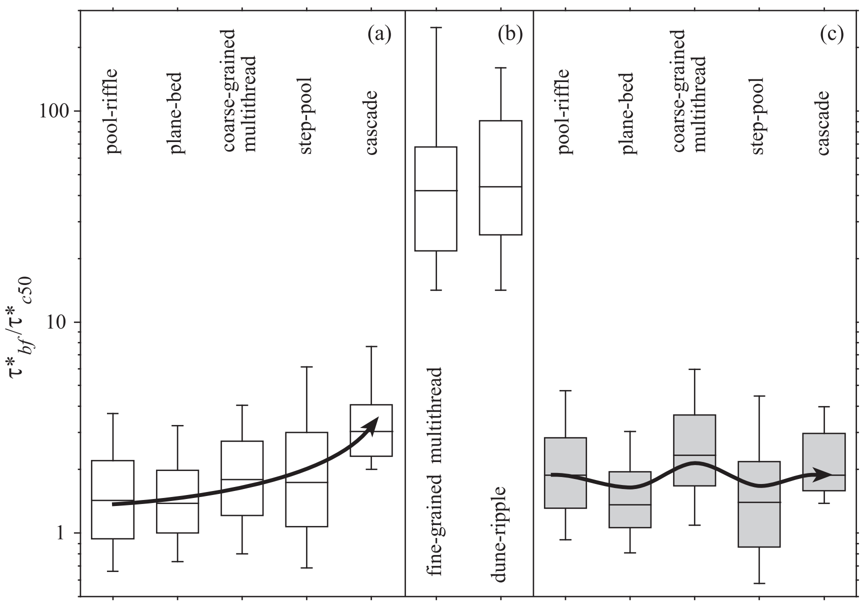

Figure 32.5 Box plots of excess bankfull Shields stress (τ^*_{bf}/τ^*_{c50}) for data of Figure 32.4. Here, multithread channels are divided into fine-grained ($D_{50} < 2$ mm, $n = 21$) and coarse-grained classes ($D_{50} \geq 2$ mm, $n = 108$). Values of τ^*_{c50} are determined from (a) Lamb *et al*. (2008), (b) Shields (1936) and (c) Pitlick *et al*. (2008).

Bed mobility indexed in terms of the excess Shields stress (τ^*_{bf}/τ^*_{c50}) shows a general increase in the bankfull mobility of D_{50} across different channel types, if the Lamb *et al*. (2008) equation is used to determine the critical Shields stress (Figure 32.5a). The systematic increase in predicted mobility is driven by the fact that the Lamb curve for τ^*_{c50} has a lower slope than the trend of the τ^*_{bf} data (Figure 32.4). In contrast, the Pitlick curve for τ^*_{c50} parallels the τ^*_{bf} data, predicting that mobility of D_{50} in coarse-grained channels is roughly constant across different channel types (median values of τ^*_{bf}/τ^*_{c50} vary from 1.4–2.3, but lack a consistent trend with channel type; Figure 32.5c). Consequently, these two τ^*_{c50} predictions represent fundamentally different interpretations of bed mobility. Field observations of mobility in different channel types suggest that the Lamb model is the more correct view; however, this issue warrants further investigation. Observed differences in mobility are discussed below in relation to Figures 32.4 and 32.5.

Dune-ripple (sand-bed) rivers are unarmoured and highly mobile, exhibiting large bankfull Shields stresses (τ^*_{bf} typically > 0.3–10; Figure 32.4) and a well-known bedform sequence that adjusts with stage and excess shear stress (Gilbert, 1914; Shields, 1936; Middleton and Southard, 1984). In contrast, pool-riffle and plane-bed channels (gravel-bed rivers) have τ^*_{bf} values that are several orders of magnitude smaller (typically > 0.01–0.4; Figure 32.4). These channels are commonly armoured and have a near-bankfull threshold for motion of the armour layer (Andrews, 1984; Buffington and Montgomery, 1999a; Ryan *et al*., 2002; 2005; Whiting and King, 2003; Mueller and Pitlick, 2005; Mueller *et al*., 2005), with relatively less mobile bedforms, except under extreme conditions (Pitlick, 1992). A near-bankfull threshold is supported by low values of τ^*_{bf}/τ^*_{c50} (median of 1.4, Figure 32.5a), while sand-bed channels are highly mobile at bankfull (median $\tau^*_{bf}/\tau^*_{c50} = 44$, Figure 32.5b). In both sand- and gravel-bed rivers, the effective discharge for sediment transport occurs frequently, corresponding with near-bankfull flows in supply-limited systems (Wolman and Miller, 1960; Andrews, 1980; Carling, 1988b; Andrews and Nankervis, 1995; Whiting *et al*., 1999; Emmett and Wolman, 2001; Crowder and Knapp, 2004; Torizzo and Pitlick, 2004) and with lesser floods in transport-limited rivers (Gomez *et al*., 2007; Ma *et al*., 2010). For gravel-bed rivers, the effective discharge and its recurrence interval vary with supply-related armouring of the bed and the resultant slope of the sediment rating curve (Emmett and Wolman, 2001; Barry *et al*., 2004; Bathurst, 2007). For both sand- and gravel-bed rivers, the effective discharge is a channel maintenance flow, with floods in excess of bankfull required for significant morphologic adjustment, particularly in resistant-boundary rivers (Pickup and Warner, 1976; Nolan *et al*., 1987;

Carling, 1988b; Gomez *et al*., 2007). Consequently, multiple levels of geomorphic effectiveness may occur. This is particularly evident in steeper-gradient channels (step-pool, cascade), where movement of boulder-sized structural elements typically requires 20–100^{+}-year floods (Hayward, 1980; Grant *et al*., 1990; Chin, 1998; Lenzi, 2001; Phillips, 2002; Lenzi *et al*., 2006b; Turowski *et al*., 2009), while smaller material is mobile annually (Schmidt and Ergenzinger, 1992; Gintz *et al*., 1996; Lenzi, 2004; Lamarre and Roy, 2008; but see Thompson and Croke, 2008). Greater mobility of D_{50} in step-pool and cascade channels is indicated by large values of both τ^*_{bf} (typically > 0.03–1; Figure 32.4) and τ^*_{bf}/τ^*_{c50} (median values of 1.7 and 3, respectively; Figure 32.5a), but these values do not represent the responsiveness of the boulder-sized structural elements of the channel.

The above observations suggest that, while systematic differences in bed mobility occur across different channel types, the mobility of a given grain size (e.g., D_{50}) may not be a good indicator of response potential in general. Size-selective transport and multiple levels of geomorphic effectiveness must also be considered, both for sediment transport and larger-scale morphologic adjustment, particularly for boulder-bed channels (Hassan and Zimmermann, Chapter 33, this volume).

Use of excess Shields stress (τ^*_{bf}/τ^*_{c50}) to interpret bed mobility also requires consideration of sediment supply and channel roughness. High bedload supply will drive textural fining and higher τ^*_{bf} values (Dietrich *et al*., 1989; Buffington and Montgomery, 1999b). Similarly, channel roughness (form drag and other losses of momentum) will cause apparent increases in τ^*_{bf} and τ^*_{c50} (Buffington and Montgomery, 1997, 2001); in Figures 32.4 and 32.5, both τ^*_{bf} and τ^*_{c50} are based on the total boundary shear stress uncorrected for roughness. Correcting boundary shear stress for channel roughness will reduce τ^*_{bf} and τ^*_{c50} values by 25–90%, depending on the magnitude and complexity of the roughness (Prestegaard, 1983; Shields and Gippel, 1995; Buffington, 1998; 2001; Buffington and Montgomery, 1999b; Kean and Smith, 2006a; Wilcox *et al*., 2006; Lamb *et al*., 2008). This correction is particularly important for steeper-gradient channel morphologies which exhibit systematic declines in: (i) width–depth ratio (greater wall roughness) and (ii) particle submergence (greater particle form drag and spill resistance) with decreasing drainage area (Buffington and Montgomery, 2001). In step-pool channels, the spacing of steps also decreases with slope, further increasing roughness (Judd and Peterson, 1969; Whittaker, 1987; Grant *et al*., 1990; Chin, 1999b). Systematic changes in channel morphology and associated roughness explain, in part, the observed relation between Shields stress and channel slope (both for the bankfull and critical values, τ^*_{bf} and τ^*_{c50}; Figure 32.4), but changes in channel hydraulics with grain emergence at steep slopes may also be important (Armanini and Gregoretti, 2005; Mueller *et al*., 2005; Vollmer and Kleinhans, 2007; Lamb *et al*., 2008; Recking, 2009). Even with roughness correction, τ^*_{bf} and τ^*_{bf}/τ^*_{c50} values will likely show relative differences in mobility with channel type (although this hypothesis remains to be tested).

32.2.3 Changes in Channel Type

At larger spatial and temporal scales, altered environmental conditions may cause changes in reach-scale channel type (Montgomery and Buffington, 1997) and planform morphology. Potential changes in channel type can be arrayed conceptually in terms of valley slope, confinement, discharge, and sediment supply, and may be associated with specific process domains (Montgomery, 1999). Recent examples are shown in Figure 32.6. The Church (2006) framework (Figure 32.6a) shows changes in alluvial channel type and stability as a function of gradient, sediment supply, and calibre of the sediment load. Figure 32.6b is similar, but also identifies process domains for: (i) the influence of vegetation, (ii) debris flows, and (iii) the occurrence of alluvial, colluvial, and bedrock channels.

Because changes in channel type involve alteration of larger-scale morphologic features, they may take centuries or more to occur (Church, 1995; Dade and Friend, 1998; Bravard, 2010). However, case studies demonstrate that such changes can occur over shorter time scales (years to decades; Schumm, 1977; 2005; Hooke, 1996; Kondolf *et al*., 2001; Surian and Rinaldi, 2003). For example, severe storms in the 1960s following a period of logging and road construction delivered massive inputs of fine sediment to the South Fork Salmon River in central Idaho. A downstream wave of fine sediment buried gravel pool-riffle channels, converting them to sand-bed dune-ripple channels (e.g., Figure 32.7), with a gradual recovery over the course of several decades (Megahan *et al*., 1980), but with numerous sand patches and sand stripes (Lisle and Hilton, 1992) persisting to the present day. River networks in central Idaho are particularly susceptible to fine-sediment disturbances because they are underlain by the Idaho Batholith, which produces abundant grus (granitic sands and silts), highlighting the importance of local geology in structuring the types of disturbances and channel responses that may occur.

At basin scales, changes in the abundance and spatial distribution of channel types can affect the availability of aquatic, riparian and hyporheic habitats, the metapopulation dynamics of fish, and the geomorphic function of

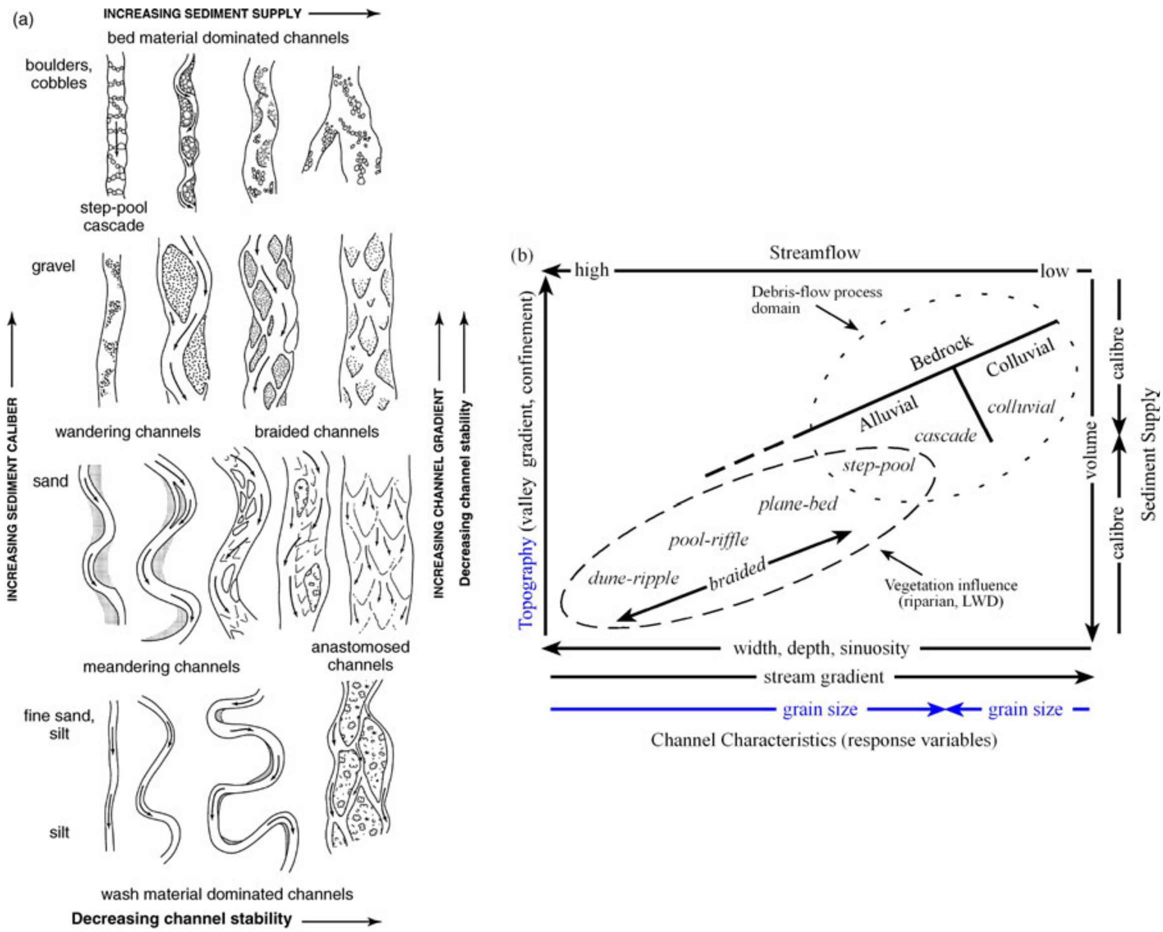

Figure 32.6 Reach-scale channel morphology as a function of imposed basin conditions (topography (gradient, confinement), discharge, sediment supply (calibre, volume)) as conceptualized by: (a) Church (2006) and (b) Buffington *et al.* (2003). Both approaches are derived from concepts developed by Mollard (1973), Brice (1975), and Schumm (1977; 1985). Panel (a) reproduced, with permission, from Church, M. 2006. Bed material transport and the morphology of alluvial rivers. Annual Review of Earth and Planetary Sciences 34, 325–354. Panel (b) modified from Buffington *et al.* (2003).

the catchment in terms of how disturbances are propagated through the river network. For example, historic effects of logging in the North American Pacific Northwest have likely altered the spatial distribution of channel types and spawning habitats for salmonids at basin scales and could have contributed to the historic decline of salmonids in this region (Buffington *et al.*, 2004a).

32.3 SPATIAL AND TEMPORAL VARIABILITY OF CHANNEL CHANGE

Probability distributions of basin inputs that drive channel response (i.e., temporal changes in water, sediment, and vegetation) vary with spatial scale. At the largest scale, physiography and climate set the sediment, hydrologic, and vegetation regimes of basins within a given region; for example, determining whether basin hydrology is dominated by snowmelt or rainfall runoff. At smaller scales within a basin, distributions of water, sediment and wood inputs change with position along the river network as a function of: (i) drainage area (Clark *et al.*, 1987; Benda and Dunne, 1997a, 1997b), (ii) process domains (Montgomery, 1999; Brardinoni and Hassan, 2006), and (iii) proximity to tributary confluences that create discontinuities or steps in basin inputs (Rice *et al.*, 2006; 2008). Finally, at the scale of stream reaches and channel units (individual pools, bars, riffles, steps), divergence of channel hydraulics and bedload transport creates patch- and point-scale probability distributions of flow and sediment supply over time

Figure 32.7 Confluence of the Middle and South Forks of the Payette River, central Idaho, showing a wave of fine sediment forcing a dune-ripple morphology (left channel) in what had previously been a gravel pool-riffle channel similar to that of the South Fork (right channel). Photo courtesy of Carter Borden. Reproduced, with permission, from Buffington, J.M., Woodsmith, R.D., Booth, D.B., Montgomery, D.R., 2003. Fluvial processes in Puget Sound Rivers and the Pacific Northwest. In: Montgomery, D.R., Bolton, S., Booth, D.B., Wall, L. (Eds.), Restoration of Puget Sound Rivers. University of Washington Press, Seattle, WA, pp. 46–78.

(seasons to years). Consequently, the driving disturbance processes and potential channel responses can be represented as a series of nested scales of probability functions that ultimately result in different disturbance–response regimes within and between basins.

Within a basin, the hydrology and sediment supply of headwater areas will be more variable than that of higher-order portions of the river network that have more stable (less variable) supplies of water and sediment due to larger contributing areas (Dunne and Leopold, 1978; Benda and Dunne, 1997a, 1997b; Church, 2002) (Figure 32.8). Furthermore, in headwater streams the magnitude of rare disturbances will be much larger per unit area due to the smaller spatial scale of headwater basins (a higher percentage of the drainage area will contribute to a given event and travel times will be shorter, with little to no opportunity for spatial attenuation of an event or for floodplain dissipation due to channel confinement; Clark *et al*., 1987; Phillips, 2002; Wohl, 2008; Goode *et al*., in press). The efficiency of disturbance propagation in headwater areas should also cause larger relative values of disturbance size (i.e., event size relative to mean annual values) and therefore should be more effective geomorphically; this effect is further magnified by channel confinement (Magilligan, 1992). In addition, floods in excess of bankfull are more frequent in headwater areas (Segura and Pitlick, 2010), further increasing potential geomorphic effectiveness. However, headwater streams have fewer degrees of freedom for channel adjustment (Table 32.1), making them resilient to all but the largest events which, when they occur, will trigger massive changes. For example, rare debris-flow events in headwater basins can cause scour to bedrock or aggradation up to tens of metres that obliterates prior channel morphology. Whether such events cause scour or deposition depends on the run-out path, which is a function of: (i) the debris-flow rheology (water–sediment content, size and sorting of sediment), (ii) channel slope, (iii) downstream changes in confinement, (iv) tributary junction angles, and (v) the presence of large wood debris (Benda and Cundy, 1990; Cenderelli and Kite, 1998; Madej, 2001; Lancaster *et al*., 2003). In contrast, lower-gradient, unconfined channels have more degrees of freedom (Table 32.1), but experience small perturbations in discharge and sediment supply relative to mean annual values because of their position in the network (larger drainage area) and because they are generally decoupled from hillslopes (Montgomery and Buffington, 1997; Church, 2002). Consequently, channel conditions are expected to be more variable over time in higher-order floodplain channels than in confined headwater portions of a basin, but the magnitude of change, when it occurs, will be larger in headwater channels (Figure 32.8). Wohl (2008) argues that flood disturbances will be most effective midway through a basin, where the relative size of an event remains moderate to large, and where the

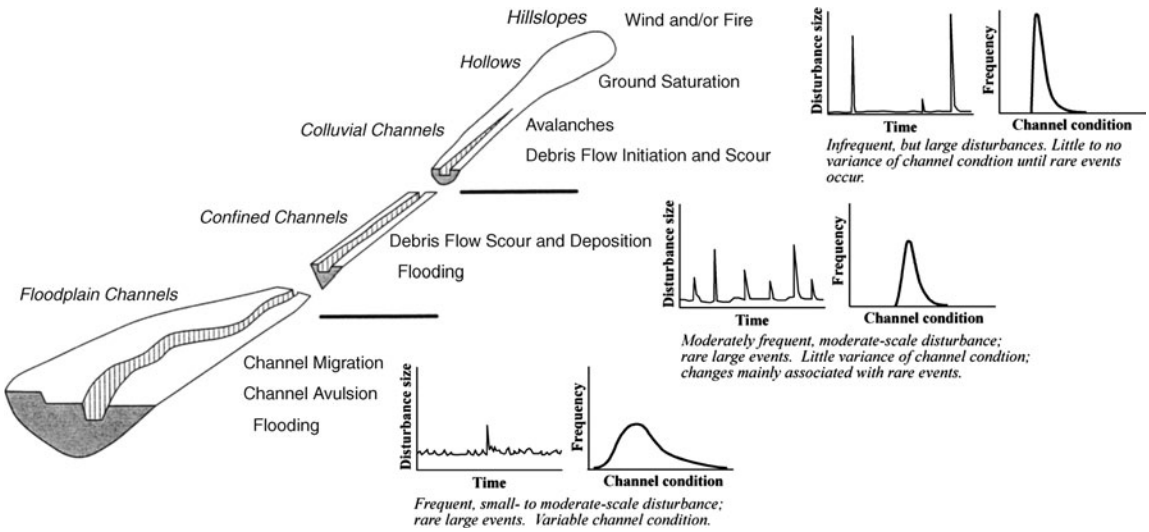

Figure 32.8 Process domains in mountain rivers, showing disturbance size relative to mean values (first right-hand graph) and variance of channel condition (second right-hand graph) as a function of these disturbances (floods, sediment inputs, changes in vegetation) and degrees of freedom associated with each process domain and channel type (Table 32.1). Modified from Montgomery (1999), and after Benda and Dunne (1997a, 1997b), Church (2002) and Wohl (2008).

more frequent occurrence of moderate-gradient floodplain channels increases the potential for channel response compared to confined, lower-order channels.

Systematic downstream changes in slope, drainage area, and process domains result in a general progression of channel types and response potentials in mountain basins (Figures 32.6 and 32.8; Schumm, 1977; 2005; Kondolf, 1994; 1995; Montgomery and Buffington, 1997; 1998; Church, 2002). However, this downstream progression can be locally re-set by a variety of factors, including changes in geology (e.g., differences in lithology or structure; Hack, 1957; Rutherford, 1994; Thompson *et al*., 2006; 2008), changes in geomorphic history (e.g., local glaciation or volcanism; Booth *et al*., 2003), and tributary confluences of high contrast (producing discontinuities in sediment, water, or wood inputs; Benda *et al*., 2004a, 2004b; Ferguson *et al*., 2006; Rice *et al*., 2006; 2008). Local geologic controls can be particularly important, causing spatial variability in hydrology (Tague *et al*., 2007; 2008), slope, rock strength, and sediment supply that produce characteristic differences in channel morphology and response potential (Buffington *et al*., 2003; Montgomery and Bolton, 2003; Brardinoni and Hassan, 2006; Thompson *et al*., 2006; 2008; Bravard, 2010).

The legacy of past natural disturbances can also affect channel morphology, network structure, and basin function, and may condition rivers for specific responses to future disturbance. In particular, the occurrence of wide valley bottoms and alluvial response reaches along the river network control the propagation of sediment pulses and floods, and their consequent downstream impacts. These features may be glacial, volcanic, structural, or climatic legacies, and are not necessarily correlated with current discharge and sediment regimes. Channel reaches may also be influenced by nested scales of legacies (geologic, climatic, infrequent events (fire, mass wasting, floods/droughts), and seasonal flows), complicating prediction of channel response to future disturbance, and making it difficult to assess causes for observed channel changes. Hence, channel response is likely complex and unique to local basin history (e.g., Carling 1988a).

32.4 PREDICTING CHANNEL CHANGE

A large body of research exists for predicting channel change and understanding morphologic stability, particularly for alluvial rivers. The goal of these studies is to predict channel characteristics (grain size, width, depth, slope, sinuosity) as a function of imposed environmental conditions (chiefly discharge and sediment supply) or, conversely, to deduce the discharge and sediment supply associated with a given channel morphology. Most of this work assumes equilibrium conditions; i.e., channel response to a given perturbation results in a stable channel form that exhibits equilibrium transport (equal rates of sediment supply and transport) at reach scales. Unfortunately, channel equilibrium has been described using a variety of terms over the years, causing some potential for confusion. For example, Lindley (1919) described unlined irrigation canals that had attained morphologic equilibrium as being "in regime", meaning the stable morphology for a given discharge and sediment transport regime (Blench 1957). As such, relations for equilibrium channel form are frequently referred to as "regime relations". In contrast, Mackin (1948) used the term "graded" to describe long-term equilibrium of natural rivers, building from observations by Gilbert (1877). The term quasi-equilibrium has also been used to describe the mean statistical state of a channel or group of channels (Blench, 1957; Langbein and Leopold, 1964). Similarly, a channel that actively moves across its floodplain, but retains long-term similarity of channel form on average, is said to be in dynamic equilibrium (Hack, 1960). Studies of channel change that invoke equilibrium conditions can be divided into two broad classes: conceptual and quantitative, each of which is examined in turn below.

32.4.1 Conceptual Models for Equilibrium Response

Most conceptual models of channel change relate equilibrium channel form to discharge and sediment supply using empirical or theoretical proportionalities. For example, Gilbert (1877) and du Boys (1879) proposed that channel slope is proportional to bedload discharge, $S \propto Q_b$. They hypothesized that an under-supplied river would erode its bed, decreasing channel slope and bedload transport rates until equilibrium transport and a metastable condition occurred. Conversely, aggradation and increased slope were expected for over-supplied rivers. Gilbert (1917) used this reasoning to explain channel response to massive sediment inputs produced by hydraulic mining in the northern California gold fields. Lane (1955b) later proposed that changes in stream discharge (Q) and bedload discharge (Q_b) are balanced by changes in channel slope (S) and grain size (D), $Q_bD \propto QS$. The Lane model is commonly visualized in terms of a scale (or balance) and is frequently invoked as a first-order conceptual model of channel response.

Work by Schumm (1977) broadened these conceptual models to include channel width (w), depth (h), planform sinuosity (p), and meander wavelength (λ) as functions of stream discharge and bedload transport rate, $Q \propto (wh\lambda)/S$, $Q_b \propto (w\lambda S)/(hp)$, with all terms in the last proportionality influenced by the silt content of the bed and bank material (Schumm, 1960a, 1960b; 1963; 1968). Schumm emphasized that channel morphology is controlled by the calibre of the sediment load and its effects on bank stability (also see Osterkamp and Hedman, 1982).

Combining the above two relations, Schumm (1977) explored potential directions of channel response to different combinations of altered discharge and sediment load (an approach further developed by Santos-Cayudo and Simons (1972), Petts (1979), Nunnally (1985), and Kellerhals and Church (1989)). These studies emphasize the potential complexity of response in floodplain alluvial channels, given their many degrees of freedom for adjustment (Table 32.1). Furthermore, they demonstrate that a given channel response (e.g., widening) can arise from multiple causes (e.g., increases in Q, Q_b, or both) and may be modulated by the response of other channel characteristics (e.g., Talbot and Lapointe, 2002a, 2002b). Consequently, it is often difficult to attribute an observed channel change to a specific cause, let alone reconstruct the exact response path, without further information. The reaction path for a given channel change depends on the sequence and magnitude of channel responses that occur during a given perturbation, as well as the history of past disturbances and channel responses to those events (i.e., legacies and pre-conditioning), as discussed above.

A second type of conceptual model involves development of evolutionary phases of channel response (i.e., "stream succession" or genetic models), building from ideas originally introduced by Davis (1889). Genetic models continue to be popular (Nanson and Croke, 1992) and are frequently used for assessing channel condition and restoration potential (Brierley and Fryirs, 2005; Rosgen, 2006). However, the Rosgen (2006) approach has been criticized because of insufficient rigor and uncertainty about using channel form to infer past processes and likely future responses (Simon *et al.*, 2007; 2008).

32.4.2 Quantitative Models for Equilibrium Response

Quantitative models of equilibrium channel response use empirical equations, analytical solutions, or numerical routines, allowing more specific predictions of response than those offered by conceptual models.

32.4.2.1 Empirical Equations

Building from work by Kennedy (1895–1896), Lindley (1919) developed empirical regime equations for the design of stable irrigation canals in the Indus River Valley, expressing channel width, depth, and slope as functions of velocity and discharge. Subsequent investigators developed a broad array of empirical regime equations for both canals and natural alluvial rivers (e.g., Lacey, 1929–1930, 1933–1934; Lane, 1937; Inglis, 1949; Blench, 1957; Simons and Albertson, 1963), including the somewhat independent development of downstream hydraulic geometry relations for natural rivers (Leopold and Maddock, 1953): $w = aQ^b$, $h = cQ^f$, $u = kQ^m$, wherein u is mean velocity, a, c, and k are empirical coefficients, and *b*, *f*, and *m* are empirical exponents. The exponents typically take values of 0.5, 0.4, and 0.1, respectively (Leopold *et al.*, 1964), but may vary regionally (Park, 1977; Wohl, 2004). Although based on different underlying empiricisms, classic regime equations can be reduced to the above hydraulic geometry relations, producing exponents similar to those above, and demonstrating the convergence of the two approaches (Leopold and Maddock, 1953; Blench, 1957; Nixon, 1959). Furthermore, stable irrigation canals and natural rivers are roughly self-similar in terms of downstream hydraulic geometry relations for mean annual discharge (Leopold and Maddock, 1953; Ferguson, 1986; Kellerhals and Church, 1989).

Hydraulic geometry relations continue to be the most popular type of empirical regime equation and have been shown to apply to a broad range of channels, including tidal marshes (Myrick and Leopold, 1963) and supraglacial streams (e.g., Marston, 1983). In mountain rivers, hydraulic geometry equations have recently been applied to bedrock rivers (Montgomery and Gran, 2001; Whipple, 2004; Finnegan *et al.*, 2005) and to headwater channels within the debris-flow process domain (Brummer and Montgomery, 2003; Buffington *et al.*, 2004b; Brardinoni and Hassan, 2007), with results reinforcing the self-similarity of channel dimensions across different channel types, and the general utility of empirical hydraulic geometry relations.

Although hydraulic geometry relations are typically created to characterize current channel morphology, they can also be used to predict changes in equilibrium channel form resulting from altered discharge (Church, 1995), or to predict past discharges from the geometry of palaeochannels (Schumm, 1968; Dury, 1976; Ethridge and Schumm, 1978; O'Connor, 1993). Similitude is commonly invoked to predict changes in channel form by expressing a given hydraulic geometry equation as a ratio of two time periods or two locations (e.g., $w_1/w_2 = a(Q_1/Q_2)^b$; Church, 1995; Griffiths, 2003); the same can be done for Lane's (1955b) proportionality (Clark and Wilcock, 2000). A limitation of this approach is that it assumes that the coefficients and exponents of the hydraulic geometry relations remain constant over time, which is unlikely in basins that undergo fundamental changes in their discharge or sediment regimes (Schumm, 1968; Ethridge and Schumm, 1978; Bray, 1982). In addition, hydraulic geometry relations produce only reconnaissance-level predictions of channel form (Dunne and Leopold, 1978); they are log-log plots with considerable uncertainty (Emmett, 1972), which may suffice for describing average changes in channel form along the downstream course of a river network, but will be inadequate to predict site-specific geometry with any degree of certainty (Buffington *et al.*, 2009). Parker (1982) argues that dimensionless hydraulic geometry relations provide more physical insight and perform better than equivalent dimensional equations (also see Parker, 1979; Andrews, 1984; Parker *et al.*, 2003; 2007), but Rhoads (1992) cautions that spurious results can result depending on the parameters selected for non-dimensionalization.

In retrospect, hydraulic geometry relations are useful first-order predictions of channel form, but they provide little mechanistic insight and, therefore, their value for advancing fluvial geomorphology is questionable (Ferguson, 1986). The self-similarity of channels across orders of magnitude in discharge is remarkable, but at the same time the value of this is suspect, given that we know that reach-scale processes and associated process domains vary along river networks. No insight about these underlying processes and controls on channel morphology are offered by hydraulic geometry relations in-and-of themselves. Eaton and Church (2007) suggest that observed downstream increases in channel width may reflect declining contributions of riparian vegetation to bank strength as the channel depth becomes large relative to typical rooting depths; for a fixed rooting depth, the vegetative cohesion is relatively high in small, shallow channels compared to large, deep rivers. However, downstream increases in channel width occur in poorly vegetated landscapes as well, indicating that vegetation is likely a modulating factor, rather than the primary control on hydraulic geometry.

32.4.2.2 Theoretical Equations

Theoretical and semi-theoretical equations for predicting channel form include a wide variety of analytical approaches and are sometimes referred to as "rational" (i.e., physically based) regime equations, particularly when written in dimensionless form (Parker, 1979; 1990; Parker *et al.*, 2003; 2007; Eaton and Church, 2007), which is another interpretation of "rational". Ideally, when developing rational solutions, one solves a system of governing equations (e.g., expressions for discharge and equilibrium bedload transport rate) that are written in terms of channel characteristics (slope, width, depth, grain size) so that the imposed watershed conditions (discharge, sediment supply) and the morphologic response variables can be solved simultaneously. The number of channel characteristics that one wishes to predict sets the number of governing equations that are needed. Results highlight that natural channels exhibit a fairly narrow range of forms compared to what is theoretically possible (rational regime equations have numerous solutions of channel form that satisfy continuity of mass and momentum, but only some of these solutions are found in nature; as recently reiterated by Eaton *et al.*, 2004). A variety of extremal hypotheses (initiated by Langbein, 1964) have been developed to explain the narrow range of observed channel forms in terms of optimal conditions (e.g., maximum sediment transporting efficiency, minimum power expenditure, etc.), which are frequently invoked as boundary conditions in solving rational regime relations (see reviews by Ferguson, 1986; Eaton *et al.*, 2004; Huang *et al.*, 2004; Millar, 2005). Extremal approaches have been criticized for their metaphysical nature (offering little process-based insight; Ferguson, 1986), their lack of agreement with observed conditions (Griffiths, 1984), and insensitivity to the particular optimal condition that is invoked (Eaton *et al.*, 2004), suggesting that in those cases the results may be mathematical artifacts of optimization, rather than having physical relevance.

The Holy Grail for rational regime equations is developing a defensible approach for predicting channel width (the most problematic channel characteristic to predict deterministically). Channel width is controlled by the erodibility of the banks which, in turn, is controlled by several variables, including: (i) the stability of cohesionless bank material (a function of grain size, friction angle, and bank slope), (ii) silt and clay content (physical cohesion), (iii) bank vegetation (root strength/biotic cohesion and roughness); (iv) bank height and degree of fluvial undercutting (risk of mass wasting), and (v) armouring by extrinsic factors (e.g., bedrock outcrops, boulders, tree roots, wood debris). Riparian vegetation is particularly important. Palaeo-reconstructions of channel morphology show that braided channels were dominant during the Precambrian to Ordovician periods, but that meandering channels became more common with the rise of terrestrial vegetation during the Late Silurian (~423–408 mya), presumably due to the additional strength and roughness of the banks provided by riparian vegetation (Montgomery *et al.*, 2003). Vegetative bank roughness affects both the magnitude and lateral distribution of boundary shear stress along the bed and banks (Houjou *et al.*, 1990; Kean and Smith, 2006a, 2006b), allowing stabilization of channel width (Parker, 1978) and promotes deposition of sands and silts during overbank flows, further stabilizing channel position and form (Smith, 2004; Allmendinger *et al.*, 2005).

Approaches for predicting channel width can be divided into three broad classes: (i) empirical equations, (ii) numerical models (ASCE, 1998; Mosselman, 1998, and Chapter 9, this volume; Nagata *et al.*, 2000; Parker *et al.*, 2011) and (iii) analytical solutions, including (a) the bankfull threshold channel (Koechlin, 1924; Glover and Florey, 1951; Lane and Carlson, 1953; Lane, 1955a; Henderson, 1963; Parker, 1978; 1979) and (b) extremal hypotheses (i.e., optimality criteria; see recent reviews by Eaton *et al.*, 2004; Huang *et al.*, 2004; Millar, 2005). None of these approaches is completely satisfying, with the pros and cons of each well discussed by Ferguson (1986) and Piégay *et al.* (2005). Parker (1978) offers mechanistic insight for the credibility of the bankfull threshold approach, but only gravel- and cobble-bed channels (pool-riffle and plane-bed morphologies) exhibit near-bankfull thresholds for mobility (Figure 32.5a), making this approach applicable only to a subset of channel types.

32.4.2.3 State Diagrams

Solutions from rational regime relations are typically presented in terms of "state diagrams" (e.g., Parker, 1990; Buffington *et al.*, 2003; Eaton *et al.*, 2004). For example, Figure 32.9a is a modified version of Parker's (1990) approach, showing unique solutions of equilibrium channel slope (S), relative submergence ($h^* = h/D_{50}$), and excess Shields stress (τ^*/τ^*_{c50}) for given pairs of dimensionless unit discharge (q^*) and bedload transport rate ($q_b{}^*$).

Here, q^* is defined from Parker *et al.* (2003; 2007) as

$$q^* = \frac{\langle u \rangle h}{\sqrt{gD_{50}}D_{50}} = \frac{\langle u \rangle h^*}{\sqrt{gD_{50}}} \tag{32.1}$$

wherein $\langle u \rangle$ is the vertically-averaged velocity determined from the law of the wall (Keulegan, 1938) as modified by Wiberg and Smith (1991) to account for particle form drag over non-uniform, rough beds

$$\langle u \rangle = \frac{u^*}{\kappa} \ln\left(\frac{0.48h}{z_0}\right) \tag{32.2}$$

In Equation (32.2), u^* is the shear velocity ($\sqrt{\tau/\rho} = \sqrt{ghS}$), κ is von Kármán's (1930) constant (0.408; Long *et al.*, 1993) and z_0 is the height above the bed where the velocity profile goes to zero ($0.1D_{84}$; Whiting and Dietrich, 1990; Wiberg and Smith, 1991), where D_{84} is the surface grain size for which 84% of the sizes are smaller). For a log-normal grain size distribution, D_{84} can be written in terms of D_{50} and the grain-size standard deviation (σ)

$$D_{84} = 2^{-(\phi_{50} - \sigma_\phi)} = D_{50} 2^{\sigma_\phi} \tag{32.3}$$

where ϕ_{50} is the median grain size in $-\log_2$ phi units (Krumbein, 1936) and σ_ϕ is the grain-size standard deviation in the same units (set equal to 1.21 ± 0.01, which is an average value for rivers with $D_{50} = 8$–256 mm; Buffington, 1999). Although shallow-flow velocity profiles are not logarithmic (Wiberg and Smith, 1991), modified forms of the law of the wall, such as Equation (32.2), can be used to estimate the *average* velocity

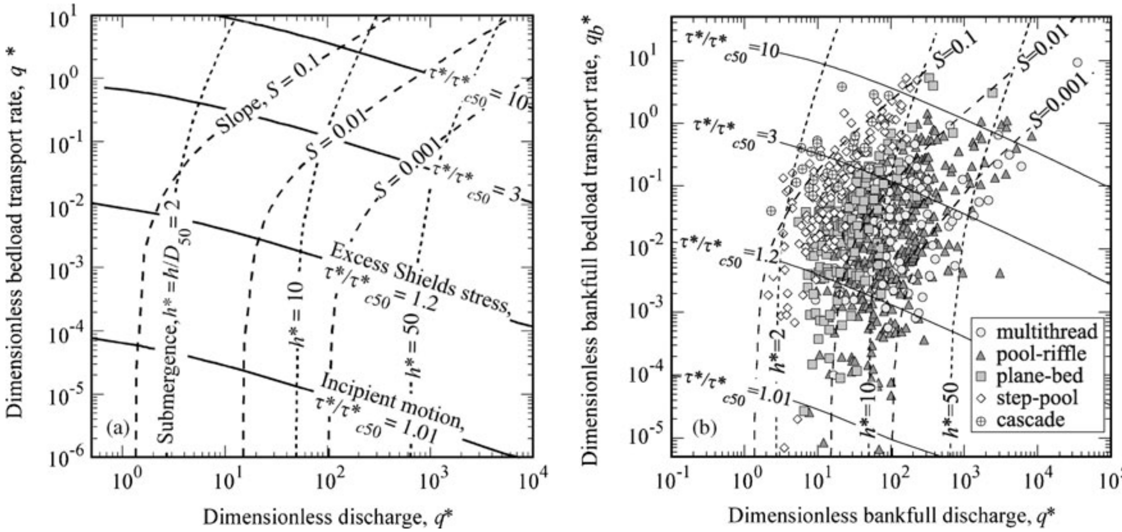

Figure 32.9 State diagram developed from Equations (32.6) and (32.7), showing: (a) contours of equilibrium channel slope (S), relative submergence ($h^* = h/D_{50}$), and excess Shields stress (τ^*/τ^*_{c50}) as functions of dimensionless discharge (q^*) and dimensionless equilibrium bedload transport rate ($q_b{}^*$), and (b) the same figure populated with field data for different reach-scale channel types evaluated at bankfull stage (data from Figure 32.4). Here, multithread channels have $D_{50} > 2$ mm. The q^* and $q_b{}^*$ values plotted in (b) are predicted from Equations (32.6) and (32.7). After Parker (1990).

of such flows; determining the full profile and the skin-friction shear stress requires numerical modelling (Wiberg and Smith, 1991). Furthermore, Equation (32.2) does not account for other sources of roughness (e.g., banks, bedforms, wood), potentially overestimating $\langle u \rangle$ in channels with multiple roughness types. Nevertheless, it is a useful first-order prediction in the absence of site-specific channel characteristics needed for more detailed partitioning of shear stress and velocity (Nelson and Smith, 1989; Shields and Gippel, 1995; Buffington, 1998; 2001; Kean and Smith, 2006a; Wilcox *et al*., 2006). Critical reviews of resistance equations and approaches for determining average velocity are presented by Ferguson (2007; in press).

The dimensionless bedload transport rate is determined here from the Meyer-Peter and Müller (1948) equation as modified by Wong and Parker (2006)

$$q_b{}^* = 4.93(\tau^* - \tau^*_{c50})^{1.6} \tag{32.4}$$

However, any suitable bedload transport equation can be used, depending on one's preference and specific goals. In Equation (32.4), the critical Shields stress is predicted as a function of S using the equation of Lamb *et al*. (2008)

$$\tau^*_{c50} = 0.15S^{0.25} \tag{32.5}$$

Equation (32.5) predicts τ^*_{c50} values based on the total boundary shear stress. To be consistent with this definition of τ^*_{c50}, the applied Shields stress in Equation (32.4) is also defined from the total boundary shear stress ($\tau^* = \rho ghS/[(\rho_s - \rho)gD_{50}] = h^*S/R$, wherein R is the submerged specific gravity, $\rho_s/\rho - 1$, with ρ_s and ρ taken as 2650 and 1000 kg m^{-3}, respectively). Hence, τ^* and τ^*_{c50} are apparent values (uncorrected for roughness). In the Meyer-Peter–Müller equation, $q_b{}^*$ is defined from the Einstein number $q_b/(\sqrt{RgD_{50}}D_{50})$, where q_b is the volumetric bedload transport rate per unit width.

Inserting the above definitions into Equations (32.1) and (32.4) yields

$$q^* = \frac{\sqrt{h^*S}h^*}{\kappa} ln\left(\frac{4.8h^*}{2^{\sigma_\phi}}\right) \tag{32.6}$$

$$q_b{}^* = 4.93(h^*S/R - \tau^*_{c50})^{1.6} = 4.93[(\tau^*/\tau^*_{c50} - 1)\tau^*_{c50}]^{1.6} \tag{32.7}$$

which expresses the dimensionless discharge (q^*) and dimensionless bedload transport rate ($q_b{}^*$) in terms of three dimensionless channel characteristics: slope (S), relative submergence ($h^* = h/D_{50}$), and excess Shields stress (τ^*/τ^*_{c50}). Analytical solutions are shown as a state diagram (Figure 32.9a), which is similar to those developed by Parker (1990) and Eaton *et al*. (2004).

The state diagram of Figure 32.9a formalizes Lane's conceptual process–response balance, offering quantitative predictions of changes in equilibrium channel form as a function of altered discharge or sediment supply ($q_b{}^*$ is a measure of both bedload transport rate and bedload sediment supply for conditions of equilibrium transport). For example, in the framework of Figure 32.9a an increase in sediment supply ($q_b{}^*$) for a constant discharge (q^*) is predicted to cause increases in channel slope (S), relative submergence (h^*), and excess Shields stress ($\tau^*/\tau^*{}_{c50}$). This can be explained as follows. The elevated sediment load initially exceeds the channel transport capacity, causing deposition and building of the bed slope, which over time increases the transport capacity of the river (Gilbert, 1877; 1917; du Boys, 1879). Increased sediment volume also leads to textural fining (greater values of both h^* and $\tau^*/\tau^*{}_{c50}$) that smoothes the bed and promotes increased transport rates (Dietrich *et al.*, 1989; Buffington and Montgomery, 1999b; Eaton and Church, 2009; Madej *et al.*, 2009). The slope and grain-size adjustments proceed until the transport capacity matches the new sediment supply. This response scenario conforms both to Lane's (1955b) conceptual framework and to field and laboratory observations of channel response (Gilbert, 1917; Madej *et al.*, 2009), but the added value of Figure 32.9a is that it allows explicit quantification of equilibrium changes in S, h^*, and $\tau^*/\tau^*{}_{c50}$.

Coupled with digital elevation models (DEMs) and predictions of reach-scale channel type as functions of stream gradient (Buffington *et al.*, 2004a), one can rapidly map the distribution of channels within a basin into state diagrams to examine potential channel responses for different disturbance scenarios (Figure 32.10). Response can be considered in terms of changes in channel characteristics (S, h^*, and $\tau^*/\tau^*{}_{c50}$), or changes in reach-scale channel type (e.g., metamorphosis from a pool-riffle channel to a plane-bed morphology). Results can then be mapped back onto the DEM to examine the spatial distribution of potential channel responses within a basin and overlain on other basin resources (e.g., spatial distributions of threatened or endangered aquatic populations and their habitat preferences) to examine the larger biophysical consequences of the predicted changes.

Solutions of equilibrium channel form obtained from state diagrams, such as Figure 32.9a, may be sensitive to how one formulates and parameterizes the governing equations (in this case, q^* and $q_b{}^*$). In particular, bedload transport equations have notoriously large potential errors if not locally calibrated (Gomez and Church, 1989; Barry *et al.*, 2004; 2007). Furthermore, state diagram

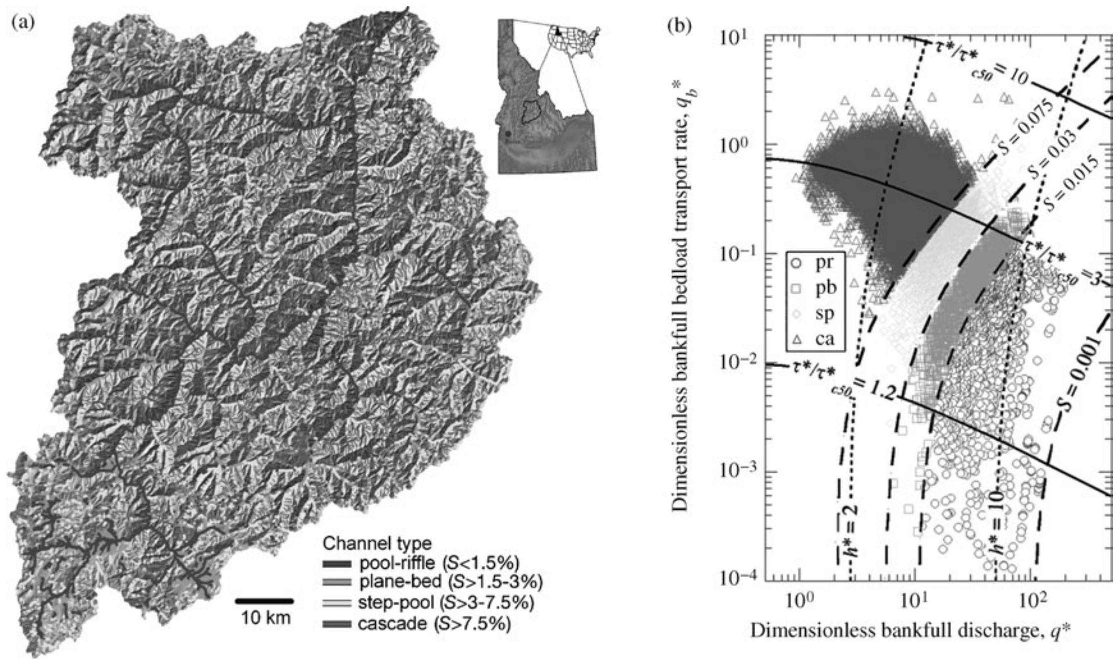

Figure 32.10 Predicted spatial distributions of: (a) reach-scale channel types in the Middle Fork Salmon River, central Idaho, USA and (b) those data mapped onto the Figure 32.9 regime diagram: pr = pool-riffle ($n = 2616$), pb = plane-bed ($n = 2056$), sp = step-pool ($n = 5983$) and ca = cascade ($n = 50\,135$). (See the color version of this figure in color plate section.)

predictions can be improved if the transport equation is tailored to channel type.

State diagrams have also been developed for predicting channel pattern (i.e., the occurrence of straight, meandering, and braided channels; Parker, 1976), but little work has been done to predict controls on other reach-scale channel types (e.g., the occurrence of pool-riffle, plane-bed, or step-pool channels), except in terms of statistical models (Wohl and Merritt, 2005; Flores *et al*., 2006; Brardinoni and Hassan, 2007; Altunkayak and Strom, 2009). Morphogenic processes are understood for some of these channel types (Montgomery and Buffington, 1997), and ranges of channel characteristics (S, w/h, h^*) have been empirically determined (Montgomery and Buffington, 1997; Chartrand and Whiting, 2000; Buffington *et al*., 2003; Wohl and Merritt, 2005; Thompson *et al*., 2006; Brardinoni and Hassan, 2007; Strom and Papanicolaou, 2009), but rational regime relations have not been developed for predicting specific reach-scale morphologies beyond the basic pattern differences discussed above. Empirically, different reach-scale channel types occupy specific domains of the Figure 32.9b state diagram highlighting two, related, well-known observations: (i) a remarkably similar suite of reach-scale channel morphologies is found across a broad range of physiographic environments and (ii) natural channels exhibit a fairly narrow range of forms compared to what is theoretically possible, with extremal hypotheses invoked to explain the latter observation, as discussed above.

One extremal hypothesis that is currently en vogue is the notion that the stable reach-scale morphology provides the roughness configuration needed for equilibrium transport in alluvial channels (i.e., channel roughness dissipates excess energy beyond that needed for equilibrium transport; Church and Jones 1982; Montgomery and Buffington, 1997; Eaton *et al*., 2004; Huang *et al*., 2004; Eaton and Church, 2007). This suggests that roughness is a key parameter in predicting channel change to altered environmental conditions and for understanding the occurrence of characteristic reach-scale morphologies. However, extremal hypotheses are theoretical constructs and limiting factors also need to be considered. For example, step-pool morphology may represent a stable roughness configuration (Whittaker and Jaeggi, 1982; Abrahams *et al*., 1995; Chin, 2002), but it requires a bimodal supply of both boulder-sized material and finer sediments, which limits the occurrence of this morphology to certain portions of the river network (i.e., steep, confined channels that are coupled to hillslope inputs of boulders, or channels that are within the debris-flow process domain, where boulders can be delivered by debris-flow events). Furthermore, oversupply of boulders will force a cascade morphology, which likely represents a jammed channel (*sensu* Church, 2006), rather than an optimal roughness response.

32.4.2.4 Empirical State Diagrams

A number of empirical state diagrams have also been developed that relate channel form to flow and sediment transport parameters, but without formal specification of the governing equations that link the two. For example, empirical state diagrams have been used to: (i) stratify phases of bed topography in sand-bed channels as functions of flow strength, depth, and grain size (Gilbert, 1914; Shields, 1936; Simons and Richardson, 1966; Allen, 1982; Middleton and Southard, 1984), (ii) examine limits on bar formation and pattern in gravel- and sand-bed rivers (Ikeda, 1975; Church and Jones, 1982; Florsheim, 1985), (iii) distinguish controls on straight, meandering, braided, and anastomosed channel patterns as a function of channel slope, discharge, grain size, and bedload transport capacity relative to sediment supply (Lane, 1957; Leopold and Wolman, 1957; Henderson, 1963; Schumm and Khan, 1972; Bray, 1982; Carson, 1984; Ferguson, 1987; Knighton and Nanson, 1993; van den Berg, 1995; Church, 2002), and (iv) delineate physical domains for different channel types (e.g., Chin, 1999a; 2002; Buffington *et al.*, 2003; Wohl and Merritt, 2005; Flores *et al.*, 2006; Brardinoni and Hassan, 2007; Church and Zimmermann, 2007; Altunkayak and Strom, 2009; Zimmermann *et al.*, 2010). The Shields curve has also been used as a state diagram to distinguish the occurrence of different channel types and modes of sediment transport as a function of excess shear stress (Shields, 1936; Dade and Friend, 1998; García, 2000; Parker *et al.*, 2003; Church, 2006; Bunte *et al.*, 2010). These empirical state diagrams can be used to predict channel change, but are not mechanistic process–response models, such as the rational regime relations discussed above.

32.4.2.5 Numerical Models

Numerical models can be used to provide quantitative predictions of channel change, but are sensitive to how the models are conceptualized, their complexity (number of dimensions, equations, and parameters), and their scale of application (Piégay *et al.*, 2005). Nevertheless, numerical models have the advantage of being able to quantify transient stages of channel change and specific response trajectories. A full discussion of the available numerical models is beyond the scope of this paper (but see recent reviews by Darby and Van de Wiel, 2003; Pizzuto, 2003; Ferguson, 2008; Mosselman, Chapter 9, this volume).

Table 32.2 Relative size of floods in different hydroclimates[a]

Discharge ratio	Snowmelt (Colorado Front Range)	Frontal rainfall (Klamath Mountains, California)[b]	Thunderstorm (Colorado Front Range)
Q_{ma}/Q_{ma}[c]	1	1	1
Q_5/Q_{ma}	1.3	1.3	1.1
Q_{10}/Q_{ma}	1.4	1.9	1.9
Q_{50}/Q_{ma}	1.8	3.5	4.5
Q_{100}/Q_{ma}	2	4.5	8.9

[a] Data from Pitlick (1994), based on regional flood frequency curves for mountain basins with roughly comparable ranges of drainage area.
[b] Approximate average value for the three frontal rainfall systems examined by Pitlick (1994).
[c] Q_{ma} = mean annual flood.

32.5 CHANNEL STABILITY AND HYDROCLIMATE

The equilibrium concept for channel adjustment is a useful first-order model and a productive means for organizing process–response theory, but it does not address channel stability (temporal variability of channel morphology). For the most part, channels are not in strict, or even quasi, equilibrium, but rather have a mean condition defined by the mode of the work function that they are exposed to (Wolman and Miller, 1960), with some degree of variance about this mean. The degree of variance (or tendency for channel change) depends on the local process domain (Sections 32.2 and 32.3). In addition, regional differences in climate can have a strong influence on basin hydrology (Pitlick, 1994; Poff *et al*., 2006), with the shape of the hydrograph likely imposing first-order controls on channel morphology and temporal variability of channel conditions (Pickup and Warner, 1976).

For example, rainfall-driven environments are characterized by flashy, peaked hydrographs, and snowmelt-driven environments are characterized by long-duration, low-amplitude hydrographs. These differences affect flood duration and the extent of morphologic adjustment that can occur. For example, ephemeral streams are typically unarmoured and lack well-defined bank and floodplain morphology due to insufficient time for bed and bank adjustments to imposed floods. In these environments, poorly developed armour layers are due to rapid recession of floods and insufficient time for selective transport and winnowing of fine grains needed for armour development (Laronne *et al*., 1994; Hassan *et al*., 2006). In contrast, snowmelt streams commonly exhibit well-armoured beds and clearly defined bankfull floodplains due to more frequent and longer-duration flood events.

Furthermore, the size of floods relative to the mean annual flow (Q_n/Q_{ma}) differs between hydroclimates, creating differences in flood variability (Table 32.2). This has several important consequences for channel morphology: (i) the more variable the flow, the more important the higher discharges become in terms of the work they accomplish relative to lower flows (Wolman and Miller, 1960; Pickup and Warner, 1976) and (ii) although the effective discharge for sediment transport occurs at bankfull flow or less in floodplain rivers, larger floods may be needed to alter channel form (Section 32.2). In particular, large floods become increasing important for lateral erosion as bank resistance increases (e.g., banks with greater vegetation, wood armouring, or large-sized sediment; Pickup and Warner, 1976; Carling, 1988b). Stevens *et al*. (1975) argue that the variance of flood size relative to the mean annual value (Table 32.2) provides an index of channel variability, with equilibrium conditions expected only in cases where Q_n/Q_{ma} does not vary substantially over time. This suggests that channel morphology may be more variable in rainfall environments than in snowmelt-dominated ones because of characteristic differences in relative flood size (Table 32.2). Furthermore, it suggests that climate-driven changes in flood variability may cause corresponding changes in the variability of channel morphology.

32.6 CONCLUSION

The potential for channel change varies over space and time as a function of the imposed environmental conditions, the available degrees of freedom for channel response, and the thresholds and time scales for a given response to occur. Consequently, channel response depends on a series of nested scales of probability functions, with unconfined alluvial channels showing the greatest potential for response. Over human time scales, rivers exhibit responses ranging from small-scale adjustments resulting from seasonal changes in watershed inputs to large-scale changes resulting from infrequent floods or decadal to centennial changes in climate (wet/dry cycles). Until recently, our ability to document these changes has been limited to: (i) broad-scale remote sensing (e.g., repeat aerial photography and satellite imagery) that typically provides fairly coarse resolution of channel features or (ii) small-scale, long-term monitoring that provides more detailed information, but only at individual cross-sections or reaches on the order of tens of channel widths in length. However, recent advances in airborne remote sensing (thermal infrared sensors (Torgersen *et al*., 2001), terrestrial and bathymetric LiDAR (McKean *et al*., 2008; 2009; Snyder, 2009; Wilkins and Snyder, 2011), and hyperspectral imagery (Marcus *et al*., 2003)) offer increasingly better resolution of data over multiple spatial scales (Marcus, Chapter 21, this volume).

Our geomorphic tool kit currently includes theoretical and empirical models for predicting channel change, but they are typically over-simplified (e.g., a preponderance of 1-D numerical models and state diagrams that are applicable to a certain class of rivers or a limited range of potential channel responses), or they are capable of providing detailed predictions only over short spatial and temporal scales (as is the case for multidimensional numerical models). Using digital elevation models to stratify the landscape by channel type and process domain offers a means to reduce the sampling effort needed and enables examination of systematic differences in geomorphic condition and response potential within and between basins (Figure 32.10; Buffington *et al*., 2004a; Benda *et al*., 2007). However, dynamic response models are needed to couple the identified process domains, and to understand system behaviour and the propagation of disturbances through a watershed.

Disturbance propagation through river networks is understood conceptually (e.g., Schumm, 1977; Montgomery and Buffington, 1997; 1998; Church, 2002), but few mechanistic models have been developed at basin scales (but see reviews by Darby and Van de Wiel (2003), Pizzuto (2003), and Ferguson (2008)). Moreover, an understanding of the legacy of past geologic, climatic, and human disturbances within a basin is essential for accurate prediction of channel response. While the above principles are generally understood, their application remains largely conceptual, and holistic models of basin function (i.e., watershed analyses) are generally lacking, hampering process-based application of river management and restoration (Beechie *et al*., 2010). In many cases, research and restoration projects are conducted in a piecemeal fashion due to limited, and generally reactive, funding (i.e., addressing perceived problems at a specific location within a basin), perpetuating a myopic focus that does little to advance our understanding of basin function.

In addition, the temporal variability of rivers is frequently under-represented. Geomorphologists and stream restoration practitioners tend to focus on mean conditions (e.g., the bankfull morphology for floodplain alluvial rivers) and to disregard channel variance, or to consider it "noise". Temporal variability is recognized, but mainly from a statistical viewpoint, rather than explicitly quantified and incorporated into our representation of fluvial processes, biophysical condition, and restoration design (i.e., designing for more than the bankfull event; Doyle *et al*., 2005; Wohl *et al*., 2005). Accounting for this variability and for non-stationarity of fluvial processes is central to assessing the potential effects of climate change on riverine ecosystems.

Finally, it is worth noting that most channel response models have been developed for alluvial, floodplain rivers, with uncertain application to steeper-gradient, confined, alluvial rivers and bedrock channels. Our understanding of hillslope–channel coupling in these environments also remains rudimentary, although some mechanistic approaches have been proposed (e.g., Whiting and Bradley, 1993; Lancaster *et al*., 2003). Consequently, further study of fluvial processes, hillslope–channel coupling and channel response potential in steep, confined rivers is warranted, particularly given that more than 80% of the river network in mountain basins is typically composed of steep, boulder-bed channels (Figure 32.10; Stock and Dietrich, 2003; Buffington *et al*., 2004a).

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32.8 REFERENCES

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32.9 APPENDIX

Data sources for Figures 32.4, 32.5, and 32.9 are: (1) Addy (2009), (2) Adenlof and Wohl (1994), (3–4) Andrews (1984; 2000), (5*) Andrews and Erman (1986), (6) Ashworth and Ferguson (1989), (**7**) BC Hydro and Power Authority (1975; 1977b; 1983), (**8**) Borland (1973), (9) Bray (1979), a subset of Kellerhals *et al.* (1972), (10) Brush (1961), (11) Buffington (unpublished data for southeast Alaska and central Idaho), (12) Buffington and Montgomery (1999), (13) Buffington *et al.* (2002; unpublished), (14) Bunte *et al.* (2010), (15) Burrows *et al.* (1981), (16) Cadol *et al.* (2009; unpublished), (<u>17</u>) Cant (1978), (18) Charlton *et al.* (1978), (19) Chitale (1970), (20) Cianfrani *et al.* (2009), (21) Clayton and Pitlick (2008), (22) Einstein (1944), (23–24) Emmett (1972; 1975), (25) Florsheim (1985), (26) Haschenburger and Church (1998), (2**7**) Galay (1971; 1977), (28) Hassan and Church (2001; unpublished), (29) Hey and Thorne (1986), (30) Higginson and Johnston (1988), (31*) Jones and Seitz (1980), (32) Judd and Peterson (1969), (**33**) Kellerhals (1967), (34) Kellerhals *et al.* (1972), (**35**) Knighton (1974), (36) Lamarre and Roy (2008), (<u>37</u>) Lambeek (1994), (38-39) Lenzi *et al.* (2006a, 2006b), (40) Leopold and Emmett (1997), (41) Leopold and Skibitzke (1967), (**42**) Lisle (1977), (43) Lisle and Madej (1992), (<u>44</u>) McCarthy *et al.* (1991), (**45**) McLean (1980), (46*) Milhous (1973), (47) Miller (1958), (48) Miller *et al.* (2002; unpublished), (49) Milner (2010; unpublished), (50) Moir *et al.* (2006), (<u>51</u>) Monsalve and Silva (1983), (52) Montgomery and Buffington (1997; unpublished), (53) Montgomery *et al.* (1995; unpublished), (54) Montgomery *et al.* (1996), (<u>55</u>) Morton and Donaldson (1978a, 1978b), (56) Mosley (1981; unpublished), (57) Mueller and Pitlick (2005), (**58**) Neil (1965), (59) Nordin and Beverage (1965), (**60**) Northwest Hydraulic Consultants (1974), (<u>61</u>) Osterkamp (1978), (62) Pitlick and Cress (2002), (63) Powell *et al.* (1998), (64) Prestegaard (1983), (<u>66</u>) Rannie (1990), (67) Roper *et al.* (2008; unpublished), (68) Ryan *et al.* (2005), (69) Sear (1993), (70*) Smalley *et al.* (1994), (71) Southerland (2003), (72) Tacconi and Billi (1987), (<u>73</u>) Taylor and Woodyer (1986), (74) Thorne and Lewin (1979), (75) Traylor and Wohl (2000), (76) Weyerhauser (unpublished), (77) Whiting *et al.* (1999), (78) Williams (1978), (<u>79</u>) Winkley (1982), (80) Wohl and Goode (2008), (81) Wohl and Wilcox (2005), (82–83) Wohl *et al.* (1993; 2004) and (**84**) Zimmerman (1975). Bold values indicate sources as cited in Church and Rood (1983), italics indicate channel pattern as defined by Church and Rood (1983), underlining indicates sources as cited by van den Berg (1995), and asterisks indicate sources as cited by Mueller *et al.* (2005).

The data sources used for each channel type are as follows: dune-ripple (**8**, 16, 19, 22–23, 30, 34–**35**, 40, <u>44</u>, <u>55</u>–56, **60**–<u>61</u>, <u>66</u>, 71, <u>73</u>, 78, <u>79</u>), pool-riffle (1, *3*, 5*–6, 9–10, 11–13, 16, 18, 20–21, 23–25, 27–29, 31*, **35**, <u>37</u>, **42**–43, 46*–50, <u>51</u>–54, 56, 62, 64, 67–70*, 71–72, 74, 76–*78*, 82–83), plane-bed (1, 4, 11–12, 14, 25, 32, 47, 49, 52–54, 57, 63, 67–68, 71, 75, 77, 81, 83), multithread (1, *7*, 9, 13, 15, <u>17</u>, *18*–19, *23*–*24*, ***26***, ***33***–34, 41, ***45***, 49, 56, ***58***–*59*, ***60***, 64, 68, *78*, ***84***,), step-pool (1–2, 11, 14, 16, 24–25, 32, 36, 38–39, 47, 49–50, 52–53, 67–68, 75–77, 80–81, 83) and cascade (1, 11, 52, 81, 83). Channel types were identified either by the original authors or from photographs, maps, and descriptions provided by a given source. Single-thread channels with $D_{50} < 2$ mm were assumed to be dune-ripple, if no other information was provided. Similarly, meandering, single-thread channels with $D_{50} > 2$ mm were assumed to be pool-riffle.

32.10 REFERENCES FOR APPENDIX

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32.11 DISCUSSION

32.11.1 Discussion by Erik Mosselman

John Buffington provides a rich bibliography on regime relations and hydraulic geometry relations, for which he rightly states channel width to be the most problematic channel characteristic to predict deterministically. However, I missed a discussion on the underlying mechanisms of width adjustment (see ASCE, 1998a). Buffington mentions bank erodibility only briefly as a control of channel width, without mechanistic details, and he does not mention bank accretion at all. Yet we know since Blench (1969) that the Holy Grail of accurate width predictors, be it empirical or rational, cannot be found if we do not consider the mechanisms of bank erosion and bank accretion, because equilibrium width is essentially multivalued and dependent on initial conditions. Moreover, the *a priori* omission of mechanisms of width adjustment produces the false idea that the channel response variables outnumber the available governing equations, and that we should hence search for some mysterious "missing equation". That this missing equation simply corresponds to the omitted equations is often forgotten by invoking extremal hypotheses instead. Objections against the latter include that extremal hypotheses do not fit in modern (Newtonian) scientific paradigms (Nagel, 1979), that they can lead to conclusions incompatible with observations (Griffiths, 1984) and that they do not make any difference with respect to contrasting hypotheses in cases where they do seem to work (Mosselman, 2004). If mechanisms of width adjustment are too difficult to include in practical applications, we may ask ourselves why we would need a single-valued width predictor in the first place. Can't we live with a range of possible widths? In stream restoration, we might either select and fix a specific width from a range of possible values (Wilcock, Chapter 12, this volume) or provide room for natural width fluctuations within certain boundaries.

32.11.2 Discussion by Jens M. Turowski

Optimization hypotheses as closure relations for regime theories of channel morphology have been debated for several decades. Supporters generally point out the good fit of such models with field data, while sceptics criticise the arbitrariness of optimal closure assumption, and the lack of direct tests for their validity. However, the physicality of such assumptions can be tested by comparing analytical regime models with steady-state solutions of dynamic channel models, when both models are based on the same physical assumptions. This was done for erosional channels by Turowski *et al*. (2009), who showed that both types of models lead to the same scaling relations between channel morphology and forcing parameters.

32.11.3 Reply by John M. Buffington

I thank Erik Mosselman for pointing out the lack of discussion regarding mechanisms for channel-width adjustment. In general, processes responsible for bank erosion include fluvial entrainment, mass wasting (frequently triggered by fluvial undercutting), and biogentic activity (e.g., tree throw and animal trampling), while channel narrowing may occur through abandonment of channel branches or bank accretion due to lateral siltation and bar growth. Thorough reviews of the processes and available models for channel-width adjustment are provided by ASCE (1998a; 1998b), Mosselman (1998), Piégay *et al*. (2005) and Rinaldi and Darby (2008).

I did not intend to suggest that width cannot be determined, or that it should be omitted from prediction of channel form. Rather, the governing equation for width is the most problematic. Some of the approaches for predicting width were discussed, but none are completely satisfying (Section 32.4.2.2). For example, uncertainty in the bank strength parameter leads to multiple width solutions in analytical approaches (Eaton *et al*., 2004). Multiple solutions are perfectly acceptable, particularly if they can be used to represent the expected natural variability of channel conditions over time. However, I would assert that we do not know the range of typical width variations for channels in general, let alone for different channel types and different hydroclimates (Section 32.5). Nor is it clear how to select width ranges in cases where channels are exhibiting transient response to natural or anthropogenic disturbances. Consequently, additional research addressing the above issues is warranted. Fixing a specific, hardened width should be discouraged, particularly for restoration, as it requires perpetual maintenance to retain that single width, which is not a responsible solution, unless required by surrounding infrastructure and human occupation of the river corridor. As suggested, designing for a range of natural width variations may be more productive (e.g., Buffington and Parker, 2005) and will avoid failure of restoration designs based on single-value (i.e., bankfull) flows, as observed in several cases studies (e.g., Kondolf *et al*., 2001; Smith and Prestegaard, 2005).

I also thank Jens Turowski for his discussion of the credibility of optimization models and for the physical insight offered by comparing the behaviour of analytical versus numerical models (Turowski *et al*., 2009).

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