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Introduction

Rivers are the shapers of terrestrial landscapes. Very few points on Earth above sea level do not lie within a drainage basin. Even points distant from the nearest channel are likely to be influenced by that channel. Tectonic uplift raises rock thousands of meters above sea level. Precipitation falling on the uplifted terrain concentrates into channels that carry sediment downward to the oceans and influence the steepness of adjacent hill slopes by governing the rate at which the landscape incises. Rivers migrate laterally across lowlands, creating a complex topography of terraces, floodplain wetlands, and channels. Subtle differences in elevation, grain size, and soil moisture across this topography control the movement of ground water and the distribution of plants and animals.

Investigators have begun to quantify the extent to which rivers influence the surrounding landscape. Stream ecologists ask, "How wide is a stream?" and address the question by using isotopic signatures to analyze food web data indicating exchanges of matter and energy between aquatic and terrestrial biotic communities (Muehlbauer et al. 2014). Geomorphologists ask, "How large is a river?" and address the question by defining signatures – emergent properties of sets of processes acting on a river landscape – and envelopes – the dynamic penetration of a signature across the landscape (Gurnell et al. 2016b). In each case, the answer is, "Wider and larger than surface appearances might suggest."

Throughout human history, people have settled disproportionately along rivers, relying on them for water supply, transport, fertile agricultural soils, waste disposal, and food from aquatic and riparian organisms. People have also devoted a tremendous amount of time and energy to altering river process and form. We are not unique in this respect: ecologists refer to various organisms, from beaver to some species of riparian trees, as *ecosystem engineers* in recognition of their ability to alter their environment. Humans are unique, however, in the extent to and intensity with which we alter rivers. In many cases, river engineering has unintended consequences, and effectively mitigating these consequences requires that we understand rivers in the broadest sense, as shapers and integrators of landscape.

Geomorphologist Luna Leopold once described rivers as the gutters down which flow the ruins of continents (Leopold et al. 1964). His father, Aldo Leopold, described the functioning of an ecosystem as a "round river," to emphasize the cycling of nutrients and energy. Rivers can be thought of as having a strong unidirectional and linear movement of water, sediment, and other materials. Rivers can also be thought of as more broadly connected systems with bidirectional fluxes of energy and matter between the channels of the river network and the greater environment. This volume emphasizes the latter viewpoint.

Rivers in the Landscape, Second Edition. Ellen Wohl.

Rivers are not simply channels. Various phrases have been used to describe the integrated system of channels, floodplain, and underlying hyporheic zone, including "the river system," "the fluvial system," "the river ecosystem," and "the river corridor." Regardless of the exact words used, the intent is to recognize that the active channel is integrally connected to adjacent surface and subsurface areas by fluxes of material and organisms. The three legs of the tripod of physical inputs that support a river corridor are inputs of water, sediment, and large wood from adjacent uplands. Although large wood has received less attention than water and sediment inputs, the historical abundance of large wood in regions with forested uplands or floodplains, along with observations of the geomorphic effects of large wood in the few remaining natural river corridors, indicates that large wood significantly influences river process and form. The material inputs of water, sediment, and wood are redistributed within the river corridor, stored for varying lengths of time, and eventually transported to the ocean, to another long-term depositional environment (e.g. alluvial fan or delta), or – for water – back to the atmosphere or ground water.

Each of the primary inputs to a river corridor can be described in terms of natural regimes that occur in the absence of human alterations in land cover, river form, flow regulation, and the water table, and in terms of altered regimes associated with human activities. The *natural flow regime* can be characterized with respect to magnitude, frequency, duration, timing, and rate of rise and fall of water discharge (Poff et al. 1997). Human alterations of the flow regime can be quantified using indicators of hydrologic alteration (Richter et al. 1996; Poff et al. 2010). The *natural sediment regime* can be characterized with respect to inputs, outputs, and storage of sediment (Wohl et al. 2015b). Because records of sediment flux analogous to those of gaged stream discharge do not exist, human alterations of the sediment regime can be inferred from the occurrence of sustained changes in river process and form that result from altered sediment dynamics. The *natural wood regime* can be characterized with respect to magnitude, frequency, duration, timing, rate, and mode of wood recruitment, transport, and storage within river corridors (Wohl et al., 2019). As with sediment, insufficient systematic records exist of wood flux in the absence of human influences to quantify changes in the natural wood regime, but the effect of human influences can be inferred from sustained changes in river process and form (e.g. Collins et al. 2012).

The details of how materials from uplands enter a river corridor and move through it are partly governed by the spatial context of the corridor (Figure 1.1). *Context* here includes valley geometry (downstream gradient, valley-bottom width relative to active channel width), position in the network, base-level stability, and substrate erosional resistance (Wohl 2018a). Valley geometry influences the energy available for changes in river form and the space available to accommodate change. Steep river reaches typically correspond to relatively narrow valleys and coarser sediment or bedrock (Livers and Wohl 2015). Lower-gradient reaches are more likely to have wide valley bottoms relative to channel width, as well as floodplains or secondary channels. Position in the network can influence the sensitivity of a river corridor to fluctuations in relative base level: commonly, the lower portions of a river network are more likely to incise in response to relative base-level fall or aggrade in response to relative base-level rise. Base-level stability influences river corridor configuration in that a river reach may be incising or aggrading irrespective of inputs of water, sediment, and large wood because of base-level instability (e.g. Schumm 1993). Substrate erosional resistance describes the ability of the channel and floodplain substrate to resist erosional changes. Resistance derives from substrate composition (grain size, stratigraphy, bedrock lithology; e.g. Finnegan et al. 2005) and from the presence of riparian vegetation (e.g. Gurnell 2014).

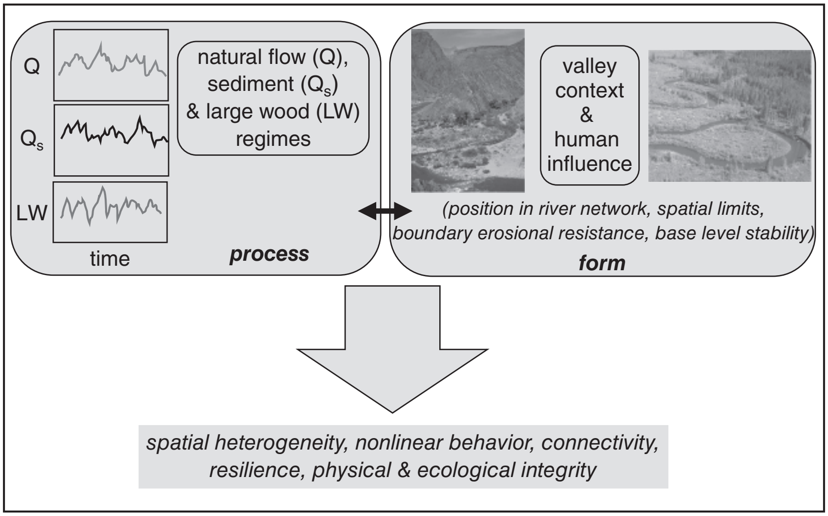

Figure 1.1 Schematic illustration of the primary inputs to river corridors (water, sediment, large wood) and the context in which they interact with one another and with the river form to create the integrative river corridor characteristics listed in the lower portion of the figure. (*See color plate section for color representation of this figure*).

Human activities can modify inputs and context. Although people typically do not alter the actual valley geometry, they do commonly alter the effective valley geometry by building levees, regulating flow and reducing flood peaks, or stabilizing the banks, each of which limits the interactions between channel and floodplain. Analogously, construction of grade controls or dams affects local base-level stability, and land drainage or bank stabilization modifies substrate erosional resistance.

Interactions between inputs and valley context create the characteristics of the river corridor listed in the lower row of Figure 1.1: spatial heterogeneity, nonlinear behavior, connectivity, resilience, and integrity. Connectivity and nonlinear behavior are introduced in this first chapter. The other concepts are covered in subsequent ones.

1.1 Connectivity and Inequality

Contemporary research and conceptual models of river form and process increasingly explicitly recognize the important of connectivity. Connectivity, sometimes referred to as coupling (e.g. Brunsden and Thornes 1979), is multifaceted. *Hydrologic connectivity* can refer to the movement of water down a hillslope in the surface or subsurface, from hillslopes into channels, or along a channel network (Pringle 2001; Bracken and Croke 2007). *River connectivity* refers to water-mediated fluxes within the channel network (Ward 1997). *Sediment connectivity* can refer to the movement, or storage, of sediment down hillslopes, into channels, or along channel networks (Harvey 1997; Fryirs et al. 2007a,b; Kuo and Brierley 2013; Bracken et al. 2015). *Biological connectivity* refers to the ability of organisms or plant propagules to disperse between suitable habitats or between isolated populations for breeding. *Landscape connectivity* can refer to the movement of water, sediment, or other materials between individual landforms such as hillslopes and channels (Brierley et al. 2006). *Structural connectivity* describes the extent to which landscape units – which can range in scale from <1 m for bunchgrasses dispersed across exposed soil to the configuration of hillslopes and valley bottoms across thousands of meters – are physically linked to one another. *Functional connectivity* describes the process-specific interactions between multiple structural characteristics, such as runoff and sediment moving downslope between the bunchgrasses and exposed soil patches (Wainwright et al. 2011). Using the scenario of runoff and sediment moving downslope, temporal variability (connectedness of rainfall) can create spatial variability (connectedness of flow paths) and thus control functional connectivity along the slope (Wainwright et al. 2011).

In general, connectivity describes the efficiency of material transfer between geomorphic system components such as floodplains and channels, hillslopes and river corridors, or longitudinal segments within a river network (Wohl et al. 2019a). Landscapes and individual landforms such as a delta are increasingly conceptualized as networks using the framework of graph and network theory (e.g. Kupfer et al. 2014; Heckmann et al. 2015; Passalacqua 2017). These networks are composed of compartments (e.g. hillslope, valley bottom), links (e.g. channel segments), and nodes (e.g. channel junctions), each of which exhibits connectivity at differing temporal and spatial scales.

Whatever form of connectivity is under discussion, its magnitude, duration, and extent are each important. Magnitude can be thought of as the volume of flux: Is only a trickle of water moving down a channel network, or a flood? Duration describes the time span of the connectivity: Can fish disperse along a river network throughout an average flow year, or only during certain seasons of high flow? Closely associated with duration is the idea of storage. If sediment stops moving downstream during periods of lower discharge, then it is at least temporarily stored in the streambed and banks. Large wood can be stored on a floodplain until overbank flows or bank erosion transport it back into the active channel. Extent is the spatial characteristic of connectivity: Does sediment move readily from the crest to the toe of a hillslope, but not into the adjacent channel because it is trapped and stored in alluvial fans perched on stream terraces? Research focuses on quantifying connectivity or developing indices of connectivity using tools such as high-resolution digital terrain models derived from aerial LiDAR (Cavalli et al. 2013) or direct measurements of fluxes (Jaeger and Olden 2012).

These dimensions of connectivity are important for adequately characterizing fluxes within a landscape, and for understanding how human activities alter those fluxes (Kondolf et al. 2006). Many human actions substantially reduce connectivity within a river network. Dams alter hydrologic connectivity and may effectively interrupt or eliminate connectivity of sediment and some organisms along a river (Magilligan et al. 2016). Levees and bank stabilization interrupt or prevent connectivity between the channel and the adjacent floodplain (Florsheim and Dettinger 2015). Flow diversions, in contrast, may increase connectivity between drainage networks, allowing exotic organisms to migrate with the diverted water and colonize a river network (Zhan et al. 2015). Dredging, channelization, straightening, and other activities that reduce geomorphic complexity and the storage of fine sediment and nutrients typically increase the longitudinal connectivity of rivers and associated downstream fluxes of sediment and solutes. By limiting overbank flows, however, these alterations reduce lateral connectivity between the channel and floodplain. Effective mitigation of undesirable human alterations of rivers requires understanding the details of connectivity.

Connectivity implies an inverse characteristic of disconnectivity. Disconnectivity can result from features that limit movement of material, typically by creating obstructions such as beaver dams (Burchsted et al. 2010) or by enhancing storage such as floodplains storing water during overbank flow (Lininger and Latrubesse 2016) or sediment (Wohl 2015a,b). Disconnectivity can also result from lack of sufficient energy or discharge to transport material in a temporally (Jaeger and Olden 2012) or spatially (Mould and Fryirs 2017) continuous manner.

Although connectivity is commonly regarded as a desirable characteristic, naturally occurring disconnectivity can be critically important. Natural disconnectivity can attenuate peak flows (Lane 2017), for example. It can also enhance retention of sediment and particulate and dissolved nutrients. This retention facilitates biological processing of these nutrients and improves water quality (Battin et al. 2008), as well as enhancing habitat abundance and diversity (Venarsky et al. 2018). A wide variety of metrics have been proposed to quantify the degree of (dis)connectivity for diverse materials (Table 1.1) (Wohl 2017b).

Table 1.1 Selected examples of quantitative metrics of connectivity.

Description	Metric	References
	Primarily hydrologic metrics	
Integral connectivity scale lengths (ICSLs)	Average distance over which wet locations are connected using either Euclidean distances or topographically defined hydrologic distances; 1 of 15 indices of hillslope hydrologic connectivity in Bracken et al. (2013: Table 4)	Western et al. (2001)
Attenuated imperviousness (I) $I = \left(\frac{\sum_j (A_j W_j)}{A_c}\right)$	Weighted impervious area as a percentage of catchment area; *Aj* is the area of the *j*th impervious surface; *Wj* is the weighting applied to *Aj*; *Ac* is catchment area	Walsh and Kunapo (2009)
River connectivity index (RCI) $DCI_P = \sum_{i=1}^{n} \frac{v_i^2}{V^2} * 100$	The size of disconnected river fragments between dams in relation to the total size of the original river network, based on Cote et al.'s (2009) directional connectivity index (DCI) model; size can be described in terms of volume (example at left), length, or other variables	Grill et al. (2014)
	Primarily sediment metrics	
Sediment delivery ratio (SDR) $SDR = \frac{net\ erosion}{total\ erosion}$	Measure of sediment connectivity	Brierley et al. (2006)
Connectivity index (IC) $IC = \log_{10}\left(\frac{D_{up}}{D_{dn}}\right)$ $D_{up} = \overline{WS}\sqrt{A}$ $D_{dn} = \sum_i \frac{d_i}{W_i S_i}$ $W = 1 - \left(\frac{RI}{RI_{MAX}}\right)$ Roughness index (RI) $RI = \sqrt{\frac{\sum_{i=1}^{25}(x_i - x_m)^2}{25}}$	D_{up} and D_{dn} are the upslope and downslope components of connectivity, respectively, with connectivity increasing as IC increases; $\overline{W}$ is the average weighting factor of the upslope contributing area, $\overline{S}$ is the average slope gradient of the upslope contributing area; A is the upslope contributing area; d_i is the length of the flow path along the ith cell according to the steepest downslope direction; W_i and S_i are the weighting factor and the slope gradient of the ith cell, respectively; RI_{MAX} is the maximum value of RI in the study area; 25 is the number of processing cells within a 5×5 moving window; x_i is the value of one specific cell of the residual topography within the moving window; x_m is the mean of the 25 cell values	Cavalli et al. (2013)

(Continued)

Table 1.1 (Continued)

Description	Metric	References
Complexity index based on overall relief Dh_{max} $Dh_{max} = E_{max} - E_{min}$ and slope variability SV $SV = S_{max} - S_{min}$	E_{max} and E_{min} are the maximum and minimum elevations, respectively, in the catchment; S_{max} and S_{min} are the maximum and minimum % slope, respectively, within the area of analysis (moving window)	Baartman et al. (2013)
Cluster persistence index (CPI) $CPI_i = \int_{\substack{over\ all \\ times\ t}} M_j^{(i)}(t)dt$	Defines clusters within a river network where mass (sediment) coalesces into a connected extent of the network; the superscript (i) denotes all clusters $M_j^{(i)}$ that occupy link i at time t	Czuba and Foufoula-Georgiou (2015)
	Metrics for diverse fluxes	
$C(t) = \sum_{i=1}^{m(t)} \sum_{j=1}^{n_i(t)} p_{ij}(t)S_{ij}(t)$	Patch connectivity, along with line, vertex, and network connectivity, can be used to characterize landscape connectivity; patch connectivity is the average movement efficiency between patches; C is patch connectivity; $p_{ij}(t)$ is the area proportion of the jth patch in the ith land cover type to the total area under investigation at time t; S is movement efficiency; $0 \leq C(t) \leq 1.1$	Yue et al. (2004)
Directional connectivity index (DCI) $DCI = \frac{\sum_{i=1}^{v} \sum_{j=r+1}^{R} w_{ij} \frac{dx(j-r)}{d_{ij}}}{\sum_{i=1}^{v} \sum_{j=r+1}^{R} w_{ij}}$	i is a node index; j is a row index; r is the row containing the node i; R is the total number of rows in the direction of interest; dx is the relative pixel length along that direction; d_{ij} is the shortest connected structural or functional distance between node i and any node in row j; w_{ij} is a weighting function	Larsen et al. (2012)
Adjacency matrix	Applies a connectivity analysis to a delta by identifying a set of objects (e.g. locations or variables) arranged in a network such that objects are nodes and connections or physical dependencies are links; connections between nodes can be evaluated using the mathematical technique of an adjacency matrix, which captures whether two nodes are connected, as well as link directionality and the strength of the connection	Newman et al. (2006); Heckmann et al. (2015); Passalacqua (2017)

Source: After Wohl (2017a,b,c), Table 2.

Inextricable from connectivity is the idea of reservoirs, sinks, or storage: components of a river channel, river network, or other landscape feature in which connectivity is at least temporarily limited. Being able to quantify the magnitude and average storage time of material in flux is critical to understanding connectivity, as is being able to predict the thresholds that define the upper and lower limits of storage. Sediment moving downslope from a weathered bedrock outcrop toward a stream channel might remain in storage on a debris-flow fan for 2000 years before reaching the stream channel, for example, so that the fan limits connectivity between the slope and channel at time spans of 10^0–10^3 years (Fryirs et al. 2007a,b). Or, the sediment might progressively accumulate on the hillslope until a precipitation or seismic trigger causes the slope to cross a threshold of stability and fail in a mass movement that instantaneously introduces much of the sediment into the stream. Or, the sediment might move quickly downslope and into the channel as soon as it is physically detached from the bedrock outcrop, because the slope angle is too high for sediment storage.

Focusing on coarse sediment transport in streams, Hooke (2003) distinguishes five scenarios. (i) Unconnected channel reaches have local sinks for sediment and lack of transport between reaches. (ii) Partially connected reaches have sediment transfer only during large floods. (iii) Connected reaches have coarse sediment transfer during frequent floods. (iv) Potentially connected reaches are competent to transfer sediment but lack a sediment supply. (v) Disconnected reaches were formerly connected but are now obstructed by a feature such as a dam. Differentiating these scenarios facilitates recognition that most natural and engineered river systems have some degree of retention of water, sediment, solutes, and organisms, and understanding net and long-term fluxes of these quantities involves quantifying both movement and storage.

Connectivity, storage, and fluxes are thus a central component of river process and form. Connectivity does not imply that all aspects of a connected valley segment, river network, or landscape are of equal importance to fluxes of matter and energy. Biogeochemists coined the phrases "hot moment" and "hot spot." A *hot moment* describes a short period of time with disproportionately high reaction rates relative to longer intervening time periods. A *hot spot* describes a small area with disproportionately high reaction rates relative to the surroundings (McClain et al. 2003). A channel-spanning logjam provides an example of a river hot spot (Figure 1.2). The logjam can effectively trap finer sediment and organic matter that might otherwise remain in transport. By storing organic matter for even a few hours, the logjam facilitates access for microbes and macroinvertebrates, which can ingest the organic matter (Bisson et al. 1987; Beckman and Wohl 2014a; Livers et al. 2018). The logjam also creates pressure gradients that facilitate downwelling of water and solutes into the streambed, where subsurface microbial communities enhance processes such as uptake of nitrate (Fanelli and Lautz 2008; Hester and Doyle 2008). The logjam thus creates a biochemical hot spot along the river.

The concepts of hot moments and hot spots are useful because any aspect of river process or form reflects inequalities in time and space. Czuba and Foufoula-Georgiou (2017), for example, identify hot spots of geomorphic change at the scale of river networks. These hot spots result from sediment accumulation or high rates of bed shear stress. Approximately 50% of the suspended sediment discharged by rivers of the Western Transverse Ranges of California, USA comes from the 10% of the basin underlain by weakly consolidated bedrock (Warrick and Mertes 2009). Somewhere between 17 and 35% of the total particulate organic carbon flux to the world's oceans comes from high-standing islands in the southwest Pacific, which constitute only about 3% of Earth's landmass (Lyons et al. 2002). Along bedrock channels with large knickpoints, the great majority of channel incision occurs at the knickpoint.

Temporal inequalities in river networks illustrate hot moments. More than 75% of the long-term sediment flux from mountain rivers in Taiwan occurs in the span of <1% of the year, during typhoon-generated floods (Kao and Milliman 2008). One-third of the total amount of stream energy generated by the Tapi River of India during the monsoon season is expended on the day of the peak flood (Kale and Hire 2007).

Figure 1.2 Channel-spanning logjam in the Rocky Mountains of Colorado, USA. Where logjams are not present, the stream has cobble- to boulder-size substrate, high transport capacity, and minimal storage of fine sediment and organic matter. Each logjam, in contrast, creates a backwater of lower-velocity flow that traps fine gravel, sand, and silt, as well as small logs, branches, and pine cones and needles. In the photograph, flow is from right to left. (*See color plate section for color representation of this figure*).

These are but a few of many examples mentioned in the remainder of this volume. Because not all moments in time or spots on a landscape are of equal importance in shaping rivers, effective understanding and management of rivers requires knowledge of how, when, and where fluxes occur.

1.2 Six Degrees of Connection

Any river network or segment of a single river exists in a rich and complicated context that reflects fluxes of matter and energy between the river and the greater environment, as well as the history of these fluxes. At any given moment in time, the only fluxes that are likely to be obvious are longitudinal fluxes as water and sediment move downstream. Longitudinal fluxes, however, are only one of six degrees of connection between a river and the environment (Figure 1.3) (Wohl 2010b).

(1) The longitudinal connection is the most obvious and intuitive. Water, sediment, and solutes move downstream. Globally, rivers transport an estimated 7819 million tons of sediment to the oceans (Milliman and Syvitski 1992) and approximately 0.9 Pg (1 Pg = 10^{15} g) of carbon per year (Aufdenkampe et al. 2011). Organisms move actively up- and downstream to new habitat and passively drift downstream with the current. Both European (*Anguilla anguilla*) and American (*Anguilla rostrata*) eels migrate from rivers to the Sargasso Sea off Bermuda for spawning, covering a

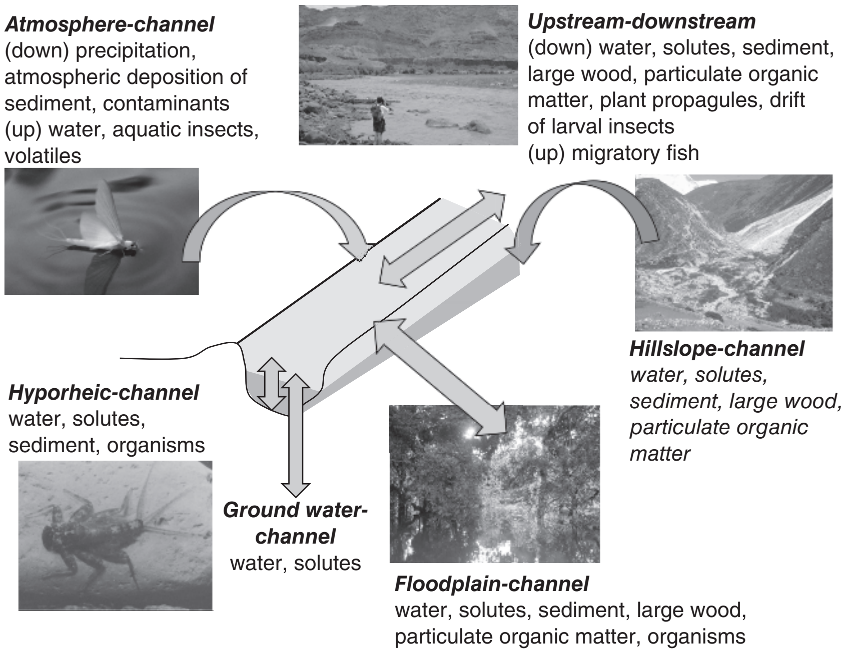

Figure 1.3 Schematic illustration of the six degrees of connection between rivers and the greater landscape. The segment of channel (lighter gray) shown here is connected to: upstream and downstream portions of the river network; adjacent uplands; the floodplain; ground water; the hyporheic zone (darker gray); and the atmosphere. The photograph representing upstream–downstream connection was taken during a flood on the Paria River, a tributary of the Colorado River that enters just downstream from Glen Canyon Dam in Arizona, USA. In this view, the Paria is turbid with suspended sediment whereas the Colorado, which is released from the base of the dam, is clear. The photograph representing hillslope–channel connection shows a large landslide entering the Dudh Khosi River in Nepal. The photograph respresenting floodplain–channel connection was taken along the Rio Jutai, a blackwater tributary of the Amazon River, during the annual flood in early June. In this view, the "flooded forest" is submerged by several meters of water. The photograph representing hyporheic–channel connection shows a larval aquatic insect (macroinvertebrate) as an example of the organisms that can move between the channel and the hyporheic environment. The photograph respresenting atmosphere–channel connection shows a mayfly emerging from the river prior to entering the atmosphere as a winged adult. Source: Image courtesy of Jeremy Monroe, Freshwaters Illustrated. (*See color plate section for color representation of this figure*).

distance of as much as 5600 km. Numerous species of salmon (*Salmo* and *Oncorhynchus* spp.) typically travel tens to hundreds of kilometers upstream from the ocean to spawn.

(2) The lateral connection between the river channel and adjacent floodplain can operate over time spans including multiple high flows as channels migrate laterally into the floodplain via bank erosion and the floodplain migrates laterally as channel bars and islands accrete to it. The lateral connection is most obvious, however, during periods of flow with sufficient volume to overtop the banks and spread across the unchanneled valley bottom. Water, sediment, solutes, and organisms disperse from the channel onto the floodplain during the rising and peak stages of a flood, and some of these materials concentrate once more in the channel during the falling stage.

High rates of primary production by photosynthetic organisms occur during the rising limb of the flood, providing food for the consumer organisms that follow the flood pulse onto the floodplain.

High rates of decomposition occur during the flood peak, and the resulting nutrients concentrate back in the channel during the descending limb. Sediment moves onto the floodplain during the rising limb, typically remaining in storage within the floodplain until bank erosion returns it to the channel (Dunne et al. 1998). Tropical river ecologists refer to the regular annual fluxes between the channel and the floodplain as the *flood pulse*, a phrase now used to refer to fluxes during floods of any recurrence interval or magnitude sufficient to create overbank flow (Junk et al. 1989; Bayley 1991). *Flow pulses* – fluctuations in surface water below the bankfull level – create similar processes within secondary channels or areas of flow separation along a single, confined channel (Tockner et al. 2000).

Levees, channelization, and flow regulation have so restricted overbank flooding along most of the world's large and medium rivers that it is now easy to underestimate the spatial extent and duration of flooding once present along lowland rivers. The Amazon, by far the world's largest river and still one of the least engineered, can extend across 50 km of floodplain during the seasonal flood, which can last more than 3 months. Along smaller rivers, historic removal of instream large wood and, in the northern hemisphere, beavers has substantially reduced channel–floodplain connectivity (Jeffries et al. 2003; Burchsted et al. 2010).

(3) The lateral connection between adjacent uplands and the river channel is more likely to be a one-way flux, with water, sediment, and solutes moving downslope at the surface and subsurface into the channel. The pathways, rates, and magnitudes of flux from the uplands typically exhibit substantial spatial and temporal variability. During an individual rainstorm, for example, water flowing across saturated ground may become a progressively more important source of runoff as infiltration capacity declines (Dykes and Thornes 2000). During the dry season, soils in the seasonal tropics can develop water repellency, which, along with an extensive network of macropores and pipes, facilitates rapid downslope transmission of runoff early in the wet season. Water repellency declines as the wet season continues, allowing infiltration to increase and runoff to decrease. By the peak of the wet season, however, saturated soils can promote rapid, abundant surface runoff (Niedzialek and Ogden 2005). Another example of temporal variability in lateral connectivity comes from rivers fed by snowmelt, which typically exhibit an ionic pulse when the release of solutes from the snowpack and the flushing of weathering products from the soil create the highest solute concentrations in the stream water at the initiation of snowmelt (Williams and Melack 1991). Mineral sediment and organic matter coming from the uplands can originate in episodic, point sources such as landslides (Hilton et al. 2008a,b) or via more diffuse, gradual erosion.

(4) Vertical fluxes link the channel to the zone of subsurface flow immediately below the channel, with flowpaths that originate and terminate at the stream. This subsurface region is known as the *hyporheic zone*, from the Greek roots *hypo* for under or beneath and *rheo* for flow or current. Water, sediment, solutes, and small organisms such as microbes and macroinvertebrates moving between the surface and subsurface can strongly influence the volume, temperature, and chemistry of flow in the river channel, and hyporheic habitat can account for a fifth of the invertebrate production in a river ecosystem (Smock et al. 1992). The hyporheic zone can extend more than 2 km laterally from the channel in wide valleys and to depths of 10 m (Stanford and Ward 1988).

(5) Deeper vertical fluxes between the river and the saturated zone of the ground water can also occur in both directions, with water and solutes moving into the channel in a *gaining stream* or into the ground water in a *losing stream*. Human activities can create gaining and losing streams. Ground-water withdrawal that lowers the water table sufficiently to prevent ground-water

flow into the channel, for example, can substantially reduce stream flow in dryland rivers (Falke et al. 2011).

As in exchanges between the hyporheic zone and surface flow, exchanges between ground and surface water can influence the temperature and chemistry of river water. Solute concentrations typically increase toward saturation as ground water moves relatively slowly through sediment or bedrock (Constantz and Stonestrom 2003), so ground-water inputs can strongly influence river solute concentrations. The flow of rivers originating from large springs in carbonate terrains or landscapes with layered basalt flows, for example, can come almost entirely from ground water (Gannett et al. 2003).

Hydraulic conductivity, a measure of permeability and ground-water flow rate, can range over 12 orders of magnitude (Domenico and Schwartz 1998). Consequently, the travel times of ground water from areas of recharge to areas of discharge in springs or rivers can range from less than a day to more than a million years (Alley et al. 2002). This means that vertical connectivity between ground water and channels typically influences river dynamics over long time scales relative to hyporheic flow.

(6) The vertical connection between the river and the atmosphere can be obvious when precipitation falls directly on the river or an aquatic insect emerges from the river for the winged, terrestrial, adult phase of its life. Other fluxes involved in this connection are likely to be much less visible. Water evaporates into the atmosphere, especially from the oceans, and moves long distances before falling onto landscapes that drain into rivers. En route, the water vapor acquires very fine particulates. These particulates include dust, which may have traveled from a different hemisphere (Prospero 1999), and nitrates from vehicles, industrial emissions, and agricultural sources. The nitrates are deposited with rain and snow – and as particles and gases – in rivers hundreds of kilometers away (Heuer et al. 2000). Fine particulates also include highly toxic mercury released by vehicles and coal-burning power plants (Grahame and Schlesinger 2007). Volatile organic compounds – solvents such as tetrachloroethylene, chlorinated compounds such as chloroform, and others – volatilize from polluted river water into the air. Although essentially invisible, these fluxes are widespread and important.

Conceptualizing a river as having six degrees of connection with the greater environment emphasizes how diverse aspects of connectivity influence river process and form. This conceptualization also emphasizes the diversity of temporal and spatial scales across which connectivity occurs. River corridor science exemplifies explicit attention to areas outside of the active channel. In hydrology, for example, the *river corridor* – the active channel(s), floodplain, riparian zone, and hyporheic zone – is an increasingly common unit of study, gradually replacing a limited focus on the wetted channel (Harvey and Gooseff 2015). In a river corridor perspective, three-dimensional exchanges and the resulting biogeochemical processing and creation and maintenance of habitat are integral to supporting healthy levels of biomass, biodiversity, water quality, and other ecosystem services associated with rivers.

1.3 Rivers as Integrators

Thanks to the extensive and sometimes subtle fluxes between a river and the greater environment, the river's forms and processes integrate the physical, chemical, and biotic processes – contemporary and historical – within the environment. This may seem obvious when considering Figure 1.3, but it represents the most profound summation possible regarding rivers, because of the implications.

If a river integrates diverse and seemingly unrelated processes within the greater environment, for example, then attempting to manage the river or some segment of the river in isolation from those processes is absurd.

If a river integrates … then human activities far from the physical boundaries of the channel may strongly influence the river, as when increasing atmospheric dust transport from the deserts of the southwestern United States alters snowpack melting and the resulting spring snowmelt hydrograph and water chemistry in rivers of the Rocky Mountains (Clow et al. 2002). Another example comes from the Mississippi River, where concentrations of nitrate have increased by two to five times since the early 1900s as farmers have applied increasing quantities of nitrogen fertilizers to upland crop fields across the Mississippi's huge drainage basin. The resulting flux of nitrate down the river to the Gulf of Mexico tripled during the last 30 years of the twentieth century, resulting in massive algal blooms that cover a swath of the Gulf as big as New Jersey (~20 000 km^2) each year, and in some years move out of the Gulf and up the eastern coast of the United States (Goolsby et al. 1999).

If a river integrates … then historical resource uses of which most people are now unaware may continue to strongly influence contemporary river process and form (Macklin and Lewin 2008). Meandering gravel-bedded streams in the eastern United States are typically bordered by fine-grained deposits that were formerly interpreted as self-formed floodplains. Prior to European settlement, however, these river networks consisted of small anabranching channels within extensive vegetated wetlands. These pre-colonial valley bottoms were buried by up to 5 m of slackwater sedimentation behind tens of thousands of seventeenth- to nineteenth-century milldams (Walter and Merritts 2008). The ubiquitous fine sediments are thus fill terraces that reflect ongoing adjustment as the milldams breached and the channels incised. Another example of historical human influences comes from rivers in the Carpathian Mountains of Poland. Agriculture began in the region during the thirteenth and fourteenth centuries, and the increased sediment yield resulted in overbank aggradation along meandering rivers draining the mountains (Klimek 1987). When the proportion of crop lands that remained bare for some portion of the year increased with more widespread cultivation of potatoes during the second half of the nineteenth century, the further increases in sediment yield caused some of the meandering rivers to assume a braided planform that persists today.

If a river integrates … then altering river process and form at one point in the river network may affect other portions of the network in unforeseen ways. The two Djerdap dams on the Danube River where it flows through Romania were built in 1970 and 1984. These massive dams, along with dozens of smaller upstream dams, have reduced sediment yields to the river's delta by 70% and silica export to the Black Sea by two-thirds relative to fluxes of these materials prior to the last third of the twentieth century. The reduced fluxes have caused erosion of the delta and a shift in the Black Sea's phytoplankton communities from siliceous diatoms to nonsiliceous coccolithophores and flagellates. These changes have stimulated algal blooms and destabilized the Black Sea ecosystem (Humborg et al. 1997). Globally, humans have increased sediment supplied to and transported by rivers as a result of soil erosion, yet reduced sediment yield to the world's oceans by 1.4 billion metric tons per year because of retention behind dams (Syvitski et al. 2005). The result of this reduced coastal sediment yield has been widespread delta and near-shore erosion (Crossland et al. 2005; Yang et al. 2011).

In summary, a river integrates fluxes across a much larger and more diverse environment than the channel itself. Consequently, understanding and effectively managing river process and form is much more challenging than is likely to be recognized if a river segment is manipulated as though it were spatially and temporally isolated.

1.4 Organization of this Volume

The title of this book, *Rivers in the Landscape*, reflects the inherent connections between a river and the landscape. Landscape is defined here as the physical, chemical, and biotic environment of the *critical zone* – Earth's outer layer, from the top of the vegetation canopy to the base of the soil and ground water, which supports life. The critical zone represents the intersection of atmosphere, water, soil, and ecosystems. Recent research increasingly reminds us of what perhaps should always have been obvious: rivers do not merely flow through a landscape in isolation, but rather interact with the landscape in complex and fascinating ways. Riverine vegetation, for example, does not just increase the hydraulic resistance of overbank flow – vegetation can alter the default river planform from braiding to meandering (Tal and Paola 2007). Rivers do not flow passively down steep topography created by tectonic uplift – removal of mass through riverine erosion can increase the upward flux of molten rock and tectonic uplift (Zeitler et al. 2001).

Recognition of the connections between rivers and landscapes implies that the topics traditionally covered in a fluvial geomorphology text – hydraulics, sediment transport, river geometry – should be treated in a manner that explicitly recognizes the influences exerted on river process and form by entities beyond the channel boundaries. Consequently, this book builds from traditional understanding of rivers toward the larger, more comprehensive viewpoint.

Chapter 2 covers the development of channels and channel networks, including how water, sediment, and solutes are produced; how they move from uplands into channels; how channel heads form; and how channel networks extend across the landscape. This chapter addresses the processes by which water moves across and through unchannelized hillslopes and concentrates sufficiently to create channels.

Chapter 3 covers channel processes, with a focus on energy (hydraulics) and quantities (hydrology). Knowledge of the basic mechanics of channelized flow is integral to understanding sediment erosion, transport and deposition, and adjustment of channel form.

Chapter 4 covers the movement of sediment in channels. The discussion begins with the sediment texture of channel beds and the processes that initiate motion of noncohesive and cohesive sediment. Once sediment is mobilized from the streambed and banks, it can be transported in solution, in suspension, or in contact with the bed, and can be organized into bedforms.

Chapter 5 discusses the movement and storage of large wood in river corridors. Starting with how wood is mobilized, transported, and deposited, the discussion explores how it influences river process and form, and the effects on rivers of human alterations of wood dynamics.

Chapter 6 addresses channel form, exploring how movement and storage of water, sediment, and large wood shape channel geometry through time and space. Interactions between process and form are implicit throughout Chapters 3–5, but Chapter 6 explicitly examines feedbacks between process and form at increasingly larger spatial scales, from cross-sectional geometry, through channel planform and longitudinal gradient, to downstream trends along a river and across a river basin.

Chapter 7 summarizes the process and form of fluvially created and maintained features outside of the active channel – floodplains, terraces, alluvial fans, deltas, and estuaries. These river landforms both reflect and influence channel process and form.

Chapter 8 metaphorically steps back to use the knowledge of process and form developed in the preceding chapters as a means to understand rivers in a landscape context. This chapter starts with a discussion of how topography influences the spatial distribution of river networks and energy expenditure within rivers, how rivers influence rates of landscape denudation, and the indicators used to infer relations between rivers and landscape evolution. Spatial differentiation of geomorphic process and form within river basins is discussed, and connectivity is reexamined. Distinctive river characteristics associated with high and low latitudes and arid regions provide examples of the importance of landscape context.

One of the challenges in writing a reasonably concise fluvial geomorphology text is the tremendous volume of research conducted on rivers within the past century. Scientists from diverse backgrounds in geology, geography, civil engineering, and other disciplines study river process and form via:

- direct measurements and experimental manipulations of real rivers;
- indirect measurements using remote sensing imagery from space-based (e.g. aerial photographs, satellite imagery, airborne LiDAR) and ground (e.g. ground-penetrating radar) platforms;
- physical experiments in a laboratory;
- numerical models; and
- integrations of these approaches.

Another fundamental challenge is the diversity of rivers. Water flows downslope under the influence of gravity. The basic physics are the same in any environment, but the ability to generalize beyond the most basic level is typically obscured by the local, place-specific details and history of a particular river. As fluvial geomorphology continues to develop as a discipline, there remains an underlying tension within the community between investigators who emphasize quantification as a means of identifying physical principles and mechanisms acting across a range of specific landscapes (e.g. Dietrich et al. 2003) and investigators who emphasize the use of historical and sedimentary records as a means of identifying the role of contingency and site-specific characteristics in river process and form (e.g. Phillips and Van Dyke 2016).

Until perhaps the 1960s or '70s, the great majority of river research focused on medium-sized, low- to medium-gradient, sand-bed rivers. These were the most accessible rivers for scientists living primarily in the temperate latitudes, and the foundational research conducted on these rivers gave rise to widely used conceptual models and equations for hydraulics, sediment transport, and channel geometry. As investigators have subsequently spent more time quantitatively examining rivers with steeper gradients and more resistant boundaries (gravel-bed rivers, bedrock rivers, mountain rivers) and greater hydrologic variability (seasonal tropics, drylands), as well as rivers at higher (boreal, arctic) and lower (tropical) latitudes, the ability of the foundational models and equations to adequately describe process and form across the known spectrum of rivers has become weaker. Throughout this volume, I explicitly address some of the unique characteristics of rivers beyond temperate-zone sand-bed channels.

My intent in this text is to maintain conciseness while reflecting the diversity of natural rivers and the methods of studying rivers. The references cited are not an exhaustive list, but rather a starting point that combines some foundational studies and particularly integrative or insightful recent studies.

1.5 Understanding Rivers

Recent emphasis on connectivity in landscapes and river networks illustrates the importance of conceptual models and methods of inquiry in governing the questions that scientists ask. If we view rivers as complex systems with multiple interactions between different components, we are more likely to focus on the factors that control those interactions and on ways to quantify and predict them. If we view rivers as predominantly physical systems, we are more likely to neglect the interactions among hydraulics, sediment dynamics, and aquatic and riparian organisms. Even when not explicitly recognized, our conceptual models of rivers tend to constrain the questions that we consider interesting and important and the methods we use to examine them (Grant et al. 2013). Studies of sediment transport, for example, that employ an *Eulerian* framework focus on the flux of sediment within a spatially bounded area. This is a very useful approach for developing a sediment budget, but a *Lagrangian* framework in which specific objects are tracked through time can provide more insight into actual mechanisms of sediment movement (Doyle and Ensign 2009).

A conceptual model results from assumptions about how a river functions. The conceptual model can be qualitative or quantitative. A quantitative model can be more precise than a qualitative model, but is not necessarily more accurate. Drawing on the second chapter of Leopold et al.'s (1964) fluvial geomorphology text for inspiration, the remainder of this section uses a landscape with which I am very familiar to explore the different conceptual models and approaches that investigators employ to understand river segments, river networks, and the greater landscape.

1.5.1 The Colorado Front Range

Atop the Precambrian-age crystalline rocks that form the continental divide in Colorado, you can stand shivering in the cold wind even at the height of summer. Here, 4000 m above sea level, bedrock topography crests in a series of ridges and peaks that divide water flowing west to the Pacific Ocean and water flowing east to the Atlantic (Figure 1.4). In some places, the divide is a sharp-edged ridge of bedrock and periglacial boulders with talus chutes and waterfalls. In other places, small alpine streams meander across broad, gently undulating surfaces.

Sharp or broad, the heights drop precipitously down to glacial cirques and troughs. Rivers alternate between paternoster lakes and steep cascades as they flow through subalpine conifer forests. Beyond the terminal glacial moraine, each valley continues downward, alternating between steep, narrow gorges in which the river flows turbulent and aerated and relatively wide canyons with gentler gradients along which the river flows through pools and riffles. These longitudinal alternations in valley and channel geometry reflect spatial heterogeneity in joint density associated with shear zones and differential weathering of the crystalline rocks. Wide, low-gradient valley segments correspond to zones with relatively densely spaced joints, whereas more widely spaced joints correspond to gorges and waterfalls (Ehlen and Wohl 2002; Wohl 2008; Ortega et al. 2013).

Climate grows progressively warmer and drier at lower elevations, and subalpine forest gives way to more open montane forest with more frequent wildfires and associated debris flows (Veblen and Donnegan 2005). Warm, moist masses of air moving inland from the southeast during summer are forced upward as they near the Colorado Rockies, and the water vapor being transported with the air masses cools, condenses, and falls as rain. Most of this moisture is wrung from the clouds at the lower to middle elevations of the mountains, which can experience flash floods from convective storms, as

Figure 1.4 Landscapes and river corridors in and adjacent to the Colorado Front Range. Upper left: View east from the summit surfaces at the continental divide. The coarse blocks in the foreground are periglacially weathered boulders and bedrock. The surface drops steeply into a glaciated valley that transitions downstream (out of sight) into a fluvial valley. Upper right: View northwest from a hogback, an asymmetrical hill of sandstone and limestone strata dipping steeply to the right in this view, with an intervening valley formed in shales. Lower right: The South Platte River near Fort Morgan, Colorado, in the low-relief environment of the Great Plain. This sand-bed channel was historically much wider and had a braided planform, but flow regulation has resulted in encroachment of riparian vegetation and transformation to a single relatively narrow channel. This river heads high in the mountains. Lower left: View of smaller drainages that head on the Great Plains, here at Pawnee National Grassland. These channels have downcut within the past few decades, largely via piping erosion. (*See color plate section for color representation of this figure*).

well as the late-spring snowmelt floods that flow down from the highest portions of the river network each year.

At the base of the mountains on the eastern side, the rivers gradually change from boulder- to cobble-bed channels as they flow through a series of steeply tilted sedimentary rocks forming asymmetrical hills. Beyond the hills lies the gently undulating topography and steppe vegetation of the semiarid Great Plains, where sand-bed channels shrink back to a trickle after the annual snowmelt peak flow.

The dramatic topography and strong elevational contrasts in climate and vegetation dominate initial impressions of the Colorado Front Range. This leads to questions about how river process and form change moving downstream, and what factors influence this change. At a basic level, we can address these questions using empirical or theoretical approaches. *Empirical* approaches are largely inductive. In logic, to induce is to conclude or infer general principles from particular examples. In an empirical approach, data are collected and analyzed in order to establish relationships between variables. A fundamental challenge to empirical understanding of rivers lies in generalizing from empirical results defined by using a restricted database. If I measure bedload transport along a cobble-bed mountain river segment for a year and demonstrate that the majority of transport occurs when flow equals or exceeds half of the bankfull depth, can I extrapolate from this site to other rivers? What if I repeat the measurements on a sand-bed river of the plains and find that bedload transport begins at a much lower level of flow?

Theoretical approaches formulate and test specific statements based on established principles. To deduce is to reason from the general to the particular. Theoretical approaches are more deductive, but are typically hampered by a relative lack of established geomorphic theory. Consequently, theoretical approaches to river process and form commonly draw heavily on related fields such as hydraulic engineering in which the theory represents a system much more simple than most natural river channels.

Theoretical approaches to bedload transport developed by hydraulic engineers, for example, assume that bedload transport (i) begins once flow energy exceeds a critical level defined by the average grain size of the sediment and (ii) is proportional to the level of excess energy beyond the critical energy. The second assumption is illustrated by a generic equation for bedload transport rate q_b

$$q_b = k(\tau - \tau_c)^n \tag{1.1}$$

where k is an empirical constant, τ is boundary shear stress, τ_c is critical boundary shear stress for entrainment, and n is an empirically derived exponent. This equation implies that bedload transport is proportional to the amount of shear stress above the critical level for moving sediment. Eq. (1.1) is an example of a *flux equation*. For rivers, flux equations usually refer to flow–sediment interactions and processes such as sediment flux within a channel.

The relatively narrow grain-size range of sand-bed channels makes it easy to specify average grain size, and the relative ease of mobility of sand grains makes assumption (ii) reasonable. In a cobble- or boulder-bed channel, however, the wider range of grain sizes means that larger grains can shield smaller grains from the force of the flow and limit the movement of the smaller grains. Consequently, average grain size may not be a particularly useful parameter for specifying the start of bedload transport. Larger grains at the streambed surface can prevent the movement of underlying smaller grains and create turbulence, so that bedload transport is not likely to have a linear relationship with flow energy.

The problem of characterizing bedload transport in mountains and plains rivers can also be described using the dichotomy of deterministic versus probabilistic. *Deterministic* approaches assume that physical laws control river process and form. Once these laws are known, river behavior can be predicted for a given set of conditions.

Deterministic modeling of river processes relies on five basic equations: one continuity equation each for water and for sediment; the flow momentum equation; a flow resistance equation; and a sediment transport equation.

Conservation equations or *continuity equations* are based on the fact that mass, momentum, and energy cannot be created or destroyed in any process. The continuity equation for flow is simply

$$Q = w\,d\,v \tag{1.2}$$

where Q is discharge, w is flow width, d is flow depth, and v is mean velocity. An example of a sediment version is the Exner equation for sediment continuity,

$$(1 - \lambda_p)\frac{\partial \eta}{\partial t} = -\frac{\partial q}{\partial x} \tag{1.3}$$

where λ_p is bed porosity, η is bed elevation, t is time, q is volume transport rate of bed material load per unit width, and x is direction of flow (Parker et al. 2000). Another example of a continuity equation is a sediment budget that equates sediment storage to sediment input minus output.

The flow momentum equation is based on Newton's second law of motion and is well defined theoretically. Momentum is a vector defined by the product of mass and velocity. Momentum per unit time of water in a channel is ρQv, where ρ is water density, Q is discharge, and v is average velocity (Robert 2014).

The flow resistance and sediment transport equations used in deterministic modeling of river processes will include empirically derived coefficients. Deterministic modeling can thus use both empirical and theoretical understanding of a system, but assumes that river process and form can be directly predicted based on knowledge of existing parameters.

As the particular component of a river system being modeled increases in complexity, the interactions are increasingly difficult to represent using a set of closed equations, and predictions become less reliable (Knighton 1998). *Probabilistic* approaches reflect an assumption that natural systems are so complex that complete deterministic explanations are unrealistic because natural systems include inherent randomness. The ability to specify appropriate empirical flow resistance and sediment transport coefficients in boulder-bed mountain streams, for example, is limited by the extreme spatial variability in bed grain size, as well as irregularities in cross-sectional geometry caused by pieces of wood and lateral constrictions from bedrock outcrops or very large boulders. Under these circumstances, it is more effective to acknowledge a substantial level of uncertainty in predicting bedload transport: bedload movement may be described as occurring when discharge falls within upper and lower bounds, rather than as a direct relationship between discharge and bedload transport (Buffington and Montgomery 1997).

Another approach to predicting bedload transport is to use a force equation. *Force equations*, typically the balance of forces involved in erosional and depositional processes, describe a critical level beyond which a process such as movement of sediment on the streambed or erosion of the stream bank begins. An example of a simple force equation for entrainment of a sediment particle in a river is

$$\tau = \gamma_f DS \tag{1.4}$$

where τ is shear stress acting on the sediment γ_f is unit weight of the fluid, D is flow depth, and S is water-surface slope (Andrews 1980). Again, the less spatial and temporal variation there is in a system – think sand-bed (relatively uniform grain sizes), rather than boulder-bed – the simpler it is to specify the forces at work and to accurately assign average values to parameters such as flow depth and water-surface slope in Eq. (1.4).

Because natural rivers are commonly quite spatially and temporally variable, geomorphologists try to simplify process and form using *physical experiments* in which one or more variables are directly and systematically manipulated in order to observe the effect on the whole system. Such manipulations are typically conducted in a laboratory setting (Schumm et al. 1987) or, more rarely, in the field.

Bedload transport in the boulder-bed mountain rivers of the Colorado Front Range occurs 24 hours a day during the snowmelt peak. Much of the transport actually occurs in the early hours of the morning when the previous afternoon's snowmelt runoff comes down the river. Instead of attempting to directly measure bedload movement, and perhaps missing some of the sediment movement by not sampling the entire channel width or sampling continuously, useful insights into sediment dynamics can be gained by creating a scaled-down river in a flume and then measuring changes in bedload transport as discharge is varied. Physical experiments present challenges of scaling forces (can you effectively simulate the turbulence and associated hydraulic forces of a flow that is several meters deep in the real channel?) and of including all relevant variables (can you effectively simulate fluctuations in upland or tributary sediment supply to the main channel?). Experiments can nonetheless provide useful insights into process and form in real channels.

Rivers can also be investigated by developing *numerical simulations* in which those variables and interactions considered to be relevant are quantified (Coulthard and Van de Wiel 2013). Simulation outcomes are then compared to real rivers in order to evaluate the accuracy of parameterization and, once such accuracy is established, to test scenarios such as the effect of altering water or sediment yields to a river. Numerical simulations can be based on some combination of theoretical and empirically derived equations, which can be deterministic or probabilistic. A numerical simulation of bedload transport, for example, might specify channel geometry, streambed grain-size distribution, discharge, and sediment input from upstream, and then use an equation such as Eq. (1.3) to predict bedload flux. Among the challenges of numerical simulation are identifying the relevant variables and processes, and parameterizing them.

In addition to downstream differences in streambed substrate and bedload dynamics, some of the more obvious changes along river networks in the Colorado Front Range are the transitions from alpine meadows to relatively dense subalpine forest of spruce and fir, to more open montane pine forest, and finally to semiarid steppe. Along the forested portions of the river networks, wood recruited from riparian forests can strongly influence channel process and form. These interactions illustrate another commonly used conceptual model of rivers as complex and nonlinear systems.

A *complex system* is composed of interconnected parts that as a whole exhibit one or more properties – including behavior – not obvious from the properties of the individual parts (Bar-Yam 1997). A complex system displays self-organization over time and emergence over scale. *Self-organization* describes the formation of patterns attributable to the internal dynamics of a system, independent of external inputs or controls (Phillips 2003). *Emergence* is defined as patterns that arise from a multiplicity of relatively simple interactions (Goldstein 1999). A tree topples into a river, for example, with a portion of the roots still attached to the bank. The downed tree extends into the river, trapping smaller pieces of wood in transport and forming a logjam. The logjam blocks flow, creating a backwater of lower velocity where sediment in transport settles out. As the elevation of the streambed increases, flow begins to spill over the channel banks and erodes a secondary channel that branches away before rejoining the main channel downstream. Bank erosion during formation of the secondary channel undermines more trees, which fall into the river, forming additional logjams that facilitate further channel branching. Eventually, a network of branching and rejoining channels that enhance wood recruitment and storage is present. The tree fall and its consequent effects resulting in a multithread channel segment are an example of a complex system (Wohl 2011b).

In a *nonlinear system*, output is not directly proportional to input, such that, mathematically, the variable to be solved for cannot be written as a linear combination of independent components because of interactions among those components (Phillips 2003). Pieces of wood floating downstream in a river are influenced by the hydraulic force of the flowing water, for example, but also by the movement of adjacent pieces of wood or the trapping effect of stationary instream wood (Braudrick et al. 1997; Kramer and Wohl 2017). Because the movement of wood down the channel does not depend only on hydraulic force, this movement is an example of a nonlinear system.

Although phrases such as “nonlinear” and “complex systems” were not commonly used until the 1990s, the behavior described by these phrases was recognized decades earlier in descriptions of river process and form within the work of G.K. Gilbert (1877) and, in the mid-twentieth century, Luna Leopold, Stanley Schumm, and others.

Rivers are also viewed as *open systems*, characterized by a continual exchange of matter and energy with the surrounding environment (Chorley 1962). Such exchanges might be obvious at the scale of a channel segment with fluxes of water and sediment *from* upstream and upland sources and *to* downstream portions of the river. As emphasized in the opening discussion of connectivity, even the largest river networks also experience continual inputs of matter and energy from the atmosphere and the lithosphere, sometimes accomplished by the activities of organisms. Snow falling in the Colorado Front Range reflects the dynamics of cold, dry Arctic air masses moving southward and interacting with slightly warmer air carrying much more moisture and moving inland from the Pacific Ocean. The melting of the resulting snowpack, and consequently the timing and volume of snowmelt runoff in the rivers, are influenced not only by air temperature, but also by deposition of wind-blown dust that can come from nearby sources such as the deserts of the southwestern United States, as well as from very distant sources in Asia (Painter et al. 2010).

Viewing a river as a complex, open system implies that, at whatever scale the river is considered, it contains multiple, interacting components. Interactions between components include feedbacks, thresholds, and lag times, and can create equifinality.

Feedback refers to interactions among variables. Self-enhancing feedbacks promote continuing change, as when sand grains saltating across bedrock are preferentially trapped on accumulations of sand that increase with time. The fallen log that initiates a logjam, and eventually a network of secondary channels that promote additional wood recruitment and logjams, is another example of a self-enhancing feedback. Self-arresting feedbacks limit change. For example, a lateral channel constriction causes an increase in flow velocity, which results in erosion of the constriction until the velocity drops below a magnitude capable of causing erosion of the channel boundaries.

Thresholds involve abrupt changes in process or form. *External or extrinsic thresholds* are crossed as a result of changes in external controls. *Internal or intrinsic thresholds* can be crossed in the absence of changes in external variables (Schumm 1979). An example of an external threshold comes from hillslope hydrology, when the early stages of precipitation cause shallow infiltration and relatively slow downslope movement of water via subsurface diffusion. When sufficient water infiltrates to reach deeper layers with preferential flow paths in the form of soil pipes, downslope water delivery to channels abruptly becomes much more rapid. In this example, the abrupt change in downslope water delivery is externally forced by increasing volumes of precipitation.

A hillslope can also be forced across a threshold of stability by intense or prolonged precipitation that saturates regolith and triggers debris flows. During widespread and sustained rainfall in September 2013, more than 1100 debris flows and landslides across the Colorado Front Range moved the equivalent of hundreds to thousands of years of hillslope weathering products (Anderson et al. 2015). Substantial amounts of large wood and organic carbon were also delivered to streams as a result of crossing the threshold of hillslope stability (Rathburn et al. 2017).

The ephemeral tributaries that head on the dry eastern steppe of Colorado provide an example of internal thresholds. Over hundreds to thousands of years, these channels alternately incised to form steep-sided gullies or arroyos and aggraded to form relatively shallow swales (Figure 1.4). These alternating episodes of cut and fill represent crossing of an external threshold in response to fluctuations in precipitation, vegetation, and runoff (Tucker et al. 2006).

Alternating cut and fill can also occur in response to the crossing of an internal threshold of sediment transport within the channel. Stream flow in such channels is brief and infrequent, and sediment can be deposited midway down as discharge declines because of evaporation and infiltration into the streambed. Repeated deposition of sediment partway along the stream's longitudinal profile can develop a steeper section of the bed, at which a headcut eventually forms, triggering a wave of upstream-migrating incision. All of this can occur in the absence of any change in external variables such as precipitation, runoff, or sediment inputs (Schumm and Hadley 1957).

Lag time typically refers to the delay between a change in an external variable, such as an increase in water yield, and the response of the river, such as bank erosion. The cobble-bed streams of the subalpine forest in the Front Range provide an example. Commercial ski resorts in this region divert river water to make snow for their ski runs. When this artificially created snow melts in spring, the runoff commonly goes into a different channel than the source of the water. These receiving channels can have peak flows more than 200% larger than would result from natural runoff. Channels along which dense riparian vegetation and cohesive silt and clay increase bank resistance take longer to respond to increased peak flows than channels with less erosionally resistant banks (David et al. 2009).

Where an external disturbance is very intense or widespread, lag times can be minimal. An intense wildfire in the montane zone of the Colorado Front Range during summer 2012 completely consumed hillslope vegetation over hundreds of hectares of pine forest underlain by weathered granite. The first rainstorms following the fire resulted in widespread erosion of hillslopes and headwater channels, and aggradation of larger channels (Wohl 2013a; Kampf et al. 2016; Rathburn et al. 2018).

Fire-induced sediment accumulation in the larger channels is of particular concern because these rivers supply municipal drinking water to communities along the base of the Front Range. Water managers trying to maintain storage capacity at intake structures and limit turbidity associated with suspended sediment and organic matter could use the force equation for sediment entrainment, the continuity equations for flow and sediment, and the flux equation for bedload transport mentioned earlier to quantify sediment transport. They could also use *diffusion equations*, which describe the movement of matter, momentum, or energy in a medium in response to some gradient, such as the turbulent mixing of suspended sediment driven by gradients in flow energy (Robert 2014). An example particularly relevant to the deposition of hillslope sediment mobilized after wildfire comes from unit sediment flux q in a river depositional system

$$q = v\frac{\partial h}{\partial x} \tag{1.5}$$

where v is diffusivity, h is elevation, and x is distance downstream (Voller and Paola 2010).

Equifinality, also known as convergence, refers to the fact that different processes and causes can produce similar effects. This condition makes it difficult to infer process from form (Chorley 1962). Channel incision leading to terrace formation along the primary rivers of the Front Range may have resulted from lowering of the base level, or from fluctuations in the water and sediment supply to the river associated with the advance and retreat of Pleistocene valley glaciers, or from widespread deforestation and mining during the nineteenth century (Schumm 1991). Data on the age, spatial extent, and stratigraphy of the terraces, as well as independent information on the timing and nature of base-level change, glaciations, and historical land use, are necessary to explain terrace formation.

Underlying conceptualizations such as feedbacks and thresholds is one of the most widely used geomorphic conceptual models: the idea that a river can exhibit various forms of stability, or equilibrium (Gilbert 1877). *Equilibrium* typically refers to a condition with no net change, and is thus very

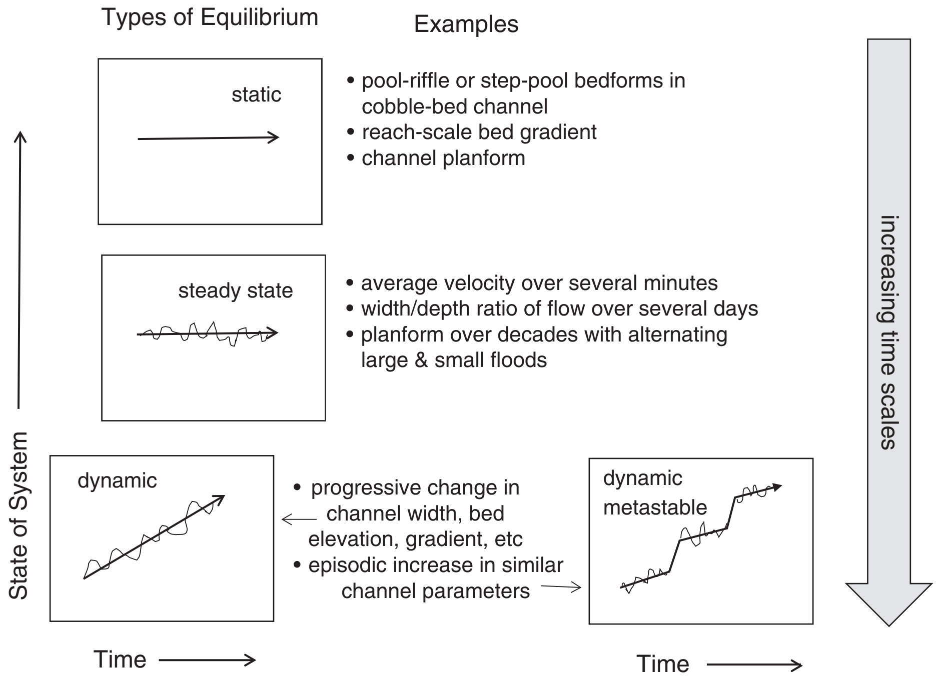

Figure 1.5 Schematic illustration of different types of equilibrium.

dependent on the time span being considered. A river that undergoes substantial channel change during a short-duration flood can nonetheless be in equilibrium when considered over a decade because subsequent smaller flows rework the erosional and depositional features created by the flood. Consequently, different forms of equilibrium can be distinguished with respect to time span (Figure 1.5).

Over the shortest time intervals, any particular variable representing the state of the river system (e.g. channel planform or gradient) is *static* and unchanging. At progressively longer time intervals, the variable may be in *steady state*, with fluctuations about a consistent mean. At the longest time intervals, the mean value of the variable is likely to change, either progressively through time in *dynamic equilibrium* or in a stepped manner that reflects the crossing of thresholds, as in *dynamic metastable equilibrium*. The latter two cases are not, strictly speaking, equilibrium, because the system exhibits net change over the time span being considered. These phrases are, however, widely used.

Equilibrium implies that multiple interacting variables within a river can reach a state of stability. This is reflected in the widely used definition of a *graded river* as a channel in which streambed slope is adjusted to prevailing water and sediment discharges, such that the channel neither aggrades nor degrades and the slope remains constant over the time interval of interest (Mackin 1948).

Equilibrium also implies that a river will change in response to changes in the supply of energy or material. Pleistocene valley glaciation in the upper portion of the Front Range changed water and sediment yields to downstream portions of the river network. Thinking of these river networks within a framework of equilibrium raises questions regarding how, and how rapidly or over what time span, the rivers responded to altered water and sediment supplies during glacial advance and retreat. One way to asses this is to evaluate downstream hydraulic geometry (Leopold and Maddock 1953) relations for rivers in the Front Range. *Downstream hydraulic geometry* relations are empirical equations in the form of *power functions* derived from linear regressions of log-transformed data. These equations relate dependent variables of channel geometry to the independent variable of discharge. For example,

$$w = aQ^b \tag{1.6}$$

where w is channel width, Q is discharge, and the coefficient a and exponent b are determined from the linear regression.

Eq. (1.6) implies that discharge is the primary influence on channel width. One implication is that values of channel width in the Front Range have fluctuated through time as the advance and retreat of valley glaciers has altered discharge downstream.

A river in equilibrium is expected to have well-developed downstream hydraulic geometry such that variations in discharge explain most of the observed downstream pattern of variation in width (Wohl 2004b). Headwater rivers within the glaciated portion of the Front Range exhibit less well-developed downstream hydraulic geometry relations, as indicated by lower values of the regression coefficient for w–Q regressions, than headwater rivers at lower elevations beyond the extent of Pleistocene glaciations. This suggests that rivers in the glaciated zone are still adjusting, more than 10 000 years after glacial retreat, to local variations in gradient, substrate resistance, sediment supply, and other factors that are affected by glaciation and can influence channel width. These rivers may be further from equilibrium than otherwise analogous channels at lower elevations in the mountain range. Downstream hydraulic geometry relations in glacial valleys, for example, include much greater variability than relations in fluvial valleys (Livers and Wohl 2015).

Equilibrium, or its absence, can also be described in terms of steady-state versus transient landscapes. A *steady-state landscape* can be defined with respect to denudation and topography as a landscape in which erosion and rock uplift are balanced such that a statistically invariant topography and constant denudation rate are maintained over a specified time interval (Whipple 2001). A steady-state landscape thus exhibits equilibrium between uplift and erosion. *Transient landscapes* are those experiencing relatively brief (on a geological time scale) increases in erosion rate in response to, for example, active tectonic uplift (Attal et al. 2008). Ongoing change indicates that a transient landscape has not yet reached equilibrium following an external perturbation.

Exhumation of the Denver Basin at the eastern margin of the Colorado Front Range within the last few million years caused relative base-level fall for the major rivers of the Front Range. Base-level fall triggered a wave of incision that has been migrating upstream at an estimated rate of 0.15 mm/yr (Anderson et al. 2006c). The location of contemporary active response to base-level fall appears as a steepening – either a waterfall or a steep section of rapids – in the longitudinal profile of each river. Portions of the river network upstream and downstream from this steeper zone are presently in steady state with respect to the base-level fall, whereas the gradient of the steeper portion of the longitudinal profile is transient.

Contrasting river process and form between different portions of a region such as the Front Range underlies another approach to understanding rivers. Data for understanding rivers can be obtained from direct measurements in a field setting or from remote sensing imagery (Oguchi et al. 2013). Because of the long temporal scales over which river processes such as development of drainage networks or longitudinal profiles act, ergodic reasoning is also commonly used. *Ergodic reasoning* substitutes space for time by comparing features in different stages of development, under the assumption that variables other than time remain relatively constant. For example, drainage networks developed on otherwise comparable basalt flows of widely differing ages within a limited region can be compared to examine network development through time. The challenge of ergodic approaches is that variables other than time likely differ between sites being compared. Even if the basalt flows are identical in composition, for example, fluctuations in climate through time might cause the older networks to represent at least preliminary development under a climate different than the climate present during development of networks on younger basalt flows.

Returning to the example of instream wood in forested streams of the Front Range, one way to investigate the importance of forest stand age to river–wood dynamics is to compare otherwise analogous stream segments flowing through forests of diverse age. Study design can be challenging: ideally, all other important parameters – drainage area, stream flow, sediment supply, valley geometry – are similar between the stream segments, and only the age of the riparian forest varies. Comparisons using this ergodic approach suggest that old-growth forests with average tree age greater than 200 years have more instream wood, larger wood pieces, more closely spaced channel-spanning logjams, and consequently a greater abundance of secondary channels and greater channel–floodplain connectivity (Livers and Wohl 2016).

Stepping back to consider the river networks of the Front Range at a regional scale, many of the questions posed by Leopold et al. (1964) remain highly relevant today:

- What factors control hillslope process and form, and the initiation of channelized flow?
  - What is the rate of bedrock weathering and regolith production?
  - How and how rapidly does regolith move downslope into channel networks?
  - What variables influence the location of channel heads?
  - What processes result in the formation of channel heads?
- What factors govern the longitudinal profile of the rivers?
  - In particular, what is the relative importance of landscape-scale denudation in response to continuing adjustment to uplift, Pleistocene glaciations, and Quaternary relative base-level fall?
  - What is the relative importance of longitudinal variations in bedrock erosional resistance, sediment supply, and flow regime?

Examining rivers in the context of the greater landscape, we can also add a series of new questions. Examples include:

- How do diverse types of connectivity vary throughout these river networks?
  - Are the alpine summit surfaces storing periglacial sediment, for example, or are they strongly coupled to adjacent glaciated valleys (Anderson et al. 2006a,b,c)?
  - Channel–floodplain and channel–hyporheic–ground water connectivity increase within lower-gradient, wider valley segments, and then decrease in steep, narrow segments. What are the specific processes governing these downstream variations in connectivity (e.g. Wegener et al. 2017)?
- What are the magnitude and extent of human alteration of river networks?
  - When people of European descent settled the Front Range during the nineteenth century, they initiated lode and placer mining, extensive deforestation, and widespread flow regulation in the form of dams and diversions (Wohl 2001). Some of these activities ceased a century ago. Do river process and form still differ between networks in which these historical activities occurred and networks that were not altered in this way?
  - How does contemporary flow regulation alter the physical and ecological functions of Front Range rivers (Ryan 1997; McCarthy 2008; Wohl and Dust 2012)?
  - Warming climate is resulting in changes in precipitation, soil moisture, wildfire regime, outbreaks of native insects that kill trees, and forest blowdowns. How do these pervasive alterations of forest dynamics and precipitation–runoff–stream flow influence channel process and form?

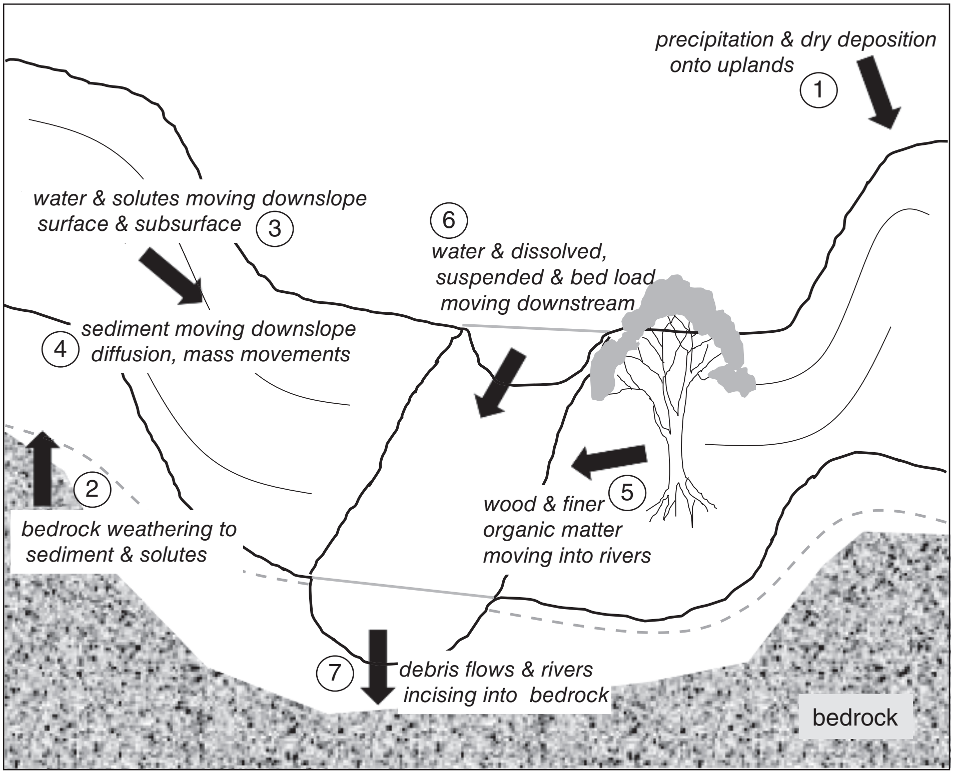

Figure 1.6 Schematic illustration of fundamental fluxes in any landscape. Among those for which some form of geomorphic transport law (GTL) has been proposed are 2, 4, and 7, although these GTLs require additional testing and parameterization for specific field settings. Empirical or theoretical equations have also been proposed for 1, 3, and 6. Again, these equations require testing and parameterization. All of the equations proposed for these fluxes assume that average values can be quantified based on prevailing conditions. Using bedrock weathering as an example, chemical reactions are a function of factors such as temperature and the amplitude of temperature oscillations. Explicitly incorporating connectivity requires quantifying variations in prevailing conditions through time that limit or enhance fluxes, such as short-term variations in weather and longer-term variations in climate that influence temperature and thus chemical weathering rate.

- What river-related geomorphic processes can be quantified in a manner applicable to diverse landscapes?
  - In an influential paper, Dietrich et al. (2003) highlighted the importance of developing geomorphic transport laws (GTLs) in the form of mathematical statements derived from a physical principle or mechanism, which express the mass flux or erosion caused by one or more processes. In order to be useful, it is important that such laws can be parameterized from field measurements, can be tested in physical models, and can be applied over relevant spatial and temporal scales. Existing GTLs include those for soil production from bedrock, linear slope-dependent transport of colluvium, and debris flow and river incision into bedrock.
  - What fundamental processes in the Colorado Front Range can usefully be expressed via GTLs (Figure 1.6)?
  - What processes are not yet adequately described by such laws?
  - How can we integrate GTLs with quantitative measures of connectivity?

- What components of river process and form are significantly influenced by biota?
  - Beaver were much more abundant in the Colorado Front Range prior to the nineteenth century. Have historical reductions in beaver populations and beaver dams influenced rivers regionally, or only local segments of rivers (Laurel and Wohl 2018)?
  - How do instream wood volume and associated geomorphic effects differ between subalpine and montane forests, or between steep, narrow valley segments and wide, lower-gradient valley segments (Livers et al. 2018)?
  - The extent and species diversity of riparian vegetation differ markedly between steep, narrow valley segments and wide, lower-gradient segments (Polvi et al. 2011). How do these differences influence valley-bottom sediment storage, hyporheic and ground water exchange, and water chemistry along Front Range rivers?

To quote Leopold et al. (1964, p. 18), "Partial explanations of these problems can be offered, but more complete explanations require much more knowledge of processes than is presently available."

1.6 Only Connect

E.M. Forster took "only connect" as the epigraph for his novel, *Howards End*. Forster was referring to connections between individual people and different classes within a society, but this phrase is also particularly apt for understanding rivers. If we can extend our understanding sufficiently and

- connect rivers to landscapes
- connect contemporary river configuration to human and geological history
- connect site-specific river characteristics to universal river process and form
- connect field observations to physical experiments and numerical simulations, and
- connect geomorphic knowledge of river process and form to
  - ecological knowledge of aquatic and riparian organisms,
  - geological knowledge of rock substrates and tectonics,
  - social science knowledge of human history and decisions regarding resource use, and
  - biogeochemical knowledge of aqueous chemistry

then we will be making progress in understanding the complex and fascinating world of rivers.