Rivers and riverine landscapes

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Introduction

The study of fluvial processes and sediment transport has a long history (e.g. Chézy, 1775; du Boys, 1879; Manning, 1891; Shields, 1936) before groundbreaking studies in the 1950s and 1960s established fundamental empirical aspects of hydraulic geometry and advanced understanding of the general processes governing river morphology and dynamics. Over the last 40 years fluvial geomorphology has grown from a focus primarily on studies of the mechanics and patterns of alluvial rivers to an expanded interest in mountain channels, the role of rivers in landscape evolution and as a geological force, and the relation of fluvial processes to aquatic and riparian ecology. An increasing emphasis on quantitative analysis and process models has forged new views of river networks as systems controlled by suites of processes, from landslide-dominated headwater valleys, to high-energy bedrock channels in mountains, lowland alluvial valleys, and estuarine channels. Key recent advances in understanding of rivers include: the processes and dynamics that lead to the development of different types of channels in different portions of a channel network; increased understanding of the fundamental coupling and interaction of rivers and tectonics; the influences of vegetation – both live and dead – on river processes and forms; and the role of riverine disturbance processes on ecological systems. In addition, advances in understanding the nature, extent, and legacies of post-glacial changes and human activities on rivers systems have increased knowledge of regional river systems. Increasingly, investigators are exploring the influences of fluvial processes on fields as diverse as the ecology of benthic macroinvertebrates and metamorphic petrology, as well as for practical efforts in conservation biology and watershed management. River restoration is emerging as an area of substantial societal investment, and presents a wealth of research opportunities in applied fluvial geomorphology.

Advances in Understanding

We cannot pretend even to attempt to review advances across the entire field of fluvial geomorphology in these few pages. Consequently, we will focus on a few topics we consider to have advanced fundamentally over the past several decades. Our review is biased and incomplete: we hope that these limitations help make it useful.

Types of Channels

Recognition that there are different types of river and stream channels is nothing new. In a paper based on his experiences with the U.S. Exploring Expedition from 1838 to 1842, J.D. Dana discussed fundamental differences between mountain channels and lowland rivers on islands of the South Pacific (Dana, 1850). Similarly, large differences in river patterns (e.g. braided, meandering, and straight) have been recognized and studied for decades. Although many fundamental aspects of river processes have been applied in the study of rivers worldwide, researchers have increasingly recognized that rivers also have distinctly regional character (e.g. rivers of the Colorado Plateau, Great Plains, Rocky Mountains, Cascades, and the coast ranges of the Pacific states). Broad variations in hydrology, geology, and vegetation impart a strong regional imprint to the morphology and dynamics of many river systems. Hydrologic regimes differ among arid, tropical, temperate, and polar regions; the geomorphic processes influencing mountain rivers differ from those in lowland regions; and the influences of vegetation reflect the dominance of forest, grassland, or shrub/scrub communities. Because different combinations of these fundamental regimes impart different characteristics to river systems in different regions, rivers are best understood in the context of their climatic and geomorphic setting, and disturbance history (Booth *et al.*, 2003; Buffington *et al.*, 2003; Montgomery, 1999; Montgomery & MacDonald, 2002).

Until recent decades research on mountain rivers and streams was eclipsed by a greater number of studies on lowland alluvial rivers. Recent work has advanced understanding of connections between process and form in mountain channel networks where reach-scale distinctions are apparent in both channel bed morphology and basin-wide relations between drainage area and slope. Montgomery & Buffington (1997) showed that different types of alluvial bed morphology in mountain channel reaches reflect the balance between transport capacity and sediment supply. Due to long-term differences in processes driving bedrock erosion, debris-flow-dominated colluvial channels and fluvial channels in upland bedrock valleys have different relations between drainage area and slope (Montgomery & Foufoula-Georgiou, 1993; Stock & Dietrich, in press). These studies showed that different portions of mountain channel networks are controlled by different processes, with key distinctions between colluvial, bedrock, and alluvial channels. In the past several decades channel and

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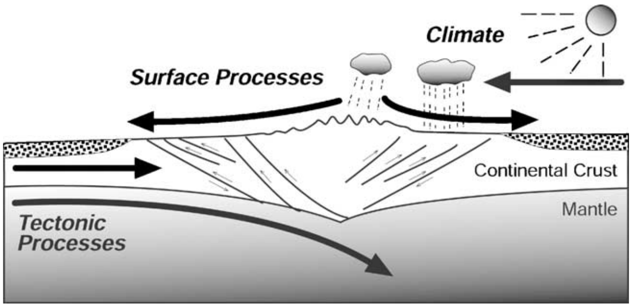

Fig. 1. Schematic illustration of relations between climate, tectonics, and erosion in shaping topography (after Willett, 1999).

floodplain classification systems have proliferated, particularly for use in the regulatory and river management arenas.

Rivers and Tectonics. Digital topography provides for quantitative, landscape-scale analyses that are modernizing the practice of geomorphology, and especially current investigations focused on relations between bedrock river incision, rock uplift, and landscape evolution (Fig. 1). The ability to analyze landforms quantitatively has been revolutionized by geographic information systems (GIS) and high-resolution topographic data. Increasing availability and resolution of digital terrain models for much of Earth's surface open new opportunities for studying the role of rivers in the evolution of particular landscapes and on interactions between rivers and tectonics. Recent interest has focused in particular on the role of rivers as a primary boundary condition on landscape evolution (Burbank *et al*., 1996; Finlayson *et al*., 2002; Montgomery & Brandon, 2002; Seidl & Dietrich, 1992; Whipple *et al*., 1999) and the role of bedload cover and sediment transport in bedrock river incision (Sklar & Dietrich, 1998, 2000). Interest in the interaction of rivers and tectonics also focuses on the role of fluvial processes in maintaining steady-state orogens (Willett & Brandon, 2002) and in the dynamics of knickpoint-dominated systems (Seidl *et al*., 1994). Research addressing spatial and temporal scales over which steady-state assumptions may be reasonable (Whipple, 2001) highlights interest in understanding the coupling of fluvial and tectonic processes.

The characteristic concave upward profiles of rivers have long been thought to reflect the downstream trade-off in erosion rates or transport capacity between increasing discharge and decreasing slope (Gilbert, 1877; Mackin, 1948). Models of river profile development predict exponential, logarithmic, or power function forms for steady-state river profiles (Snow & Slingerland, 1987), and deviations from expected trends are interpreted to reflect differences in either geologic or climate history, or spatial variability in erosion resistance, erosional processes, or rock uplift rates (Hack, 1957; Snow & Slingerland, 1990). Seeber & Gornitz (1983), for example, used the alignment of knickpoints on trans-Himalayan rivers to argue for active deformation along the Main Central Thrust between the Lesser and High Himalaya. Aided by digital elevation models (DEMs), Seidl *et al*. (1996) used the longitudinal profiles of rivers draining the flank of the south-east Australian escarpment to investigate the kinematics and pattern of escarpment retreat. Based on the spatial coincidence of distinct knickpoints at the head of major tributaries they inferred that long-term escarpment retreat at about 2 mm yr^{-1} was controlled by rock strength and fracturing more than by fluvial discharge or stream power. Hence, analyses of DEM-derived river profiles can help to elucidate the mechanisms behind the long-term evolution of both active and passive margins.

Many workers report that channel slope varies as an inverse power function of drainage area

$$S = cA^{-\theta} \tag{1}$$

where θ varies from 0.2 to 1.0 (Flint, 1974; Hack, 1957; Hurtrez *et al*., 1999; Kirby & Whipple, 2001; Moglen & Bras, 1995; Snyder *et al*., 2000; Tarboton *et al*., 1989). Headwater channels prone to debris flows exhibit different values of θ than do downstream fluvial channels (Montgomery & Foufoula-Georgiou, 1993; Seidl & Dietrich, 1992), and plots of drainage area vs. channel slope have been used to characterize different portions of a river system dominated by different processes (Montgomery, 2001; Montgomery & Foufoula-Georgiou, 1993; Snyder *et al*., 2000).

Over the past decade it has become common for the local erosion rate (E) to be modeled as a function of drainage area (A) and local slope (S) for detachment-limited channel incision

$$E = KA^m S^n, \tag{2}$$

where K is an empirical coefficient that incorporates climatic factors and bedrock erodibility, and m and n are thought to vary with different erosional processes. For the special case of steady-state topography, the local erosion rate at a distance x along the channel $E(x)$ everywhere equals the local rock uplift rate $U(x)$, and Eq. (2) can be rearranged to yield a relation between drainage area and slope

$$S = \left[\frac{U(x)}{K(x)}\right]^{1/n} A^{-(m/n)} \tag{3}$$

For spatially uniform rock uplift and lithology (i.e. $U(x)$ and $K(x)$ are constants), Eqs (1) and (3) imply that for steady-state topography $c = [U/K]^{1/n}$ and $\theta = m/n$. Models of detachment limited bedrock river incision based on both shear stress and unit stream power formulations hold that $m/n \approx 0.5$ (Whipple & Tucker, 1999).

Stock & Montgomery (1999) analyzed patterns of 13 rivers where initial river profiles of known age were compared with modern river profiles to constrain possible values of K and m/n. They found that for roughly half of the available examples the optimal m/n value ranged from 0.3 to 0.5, but that for the other half of the cases studied there was only a weak area dependence, with $m/n \approx 0.1$–0.2. They also found that K varied by at least first orders of magnitude among different lithologies, implying a huge range of potential time scales of landscape response to changes in climate or tectonic forcing. Hence, in many cases the assumption of steady state may be difficult to justify.

Lague *et al.* (2000) rearranged Eq. (3) to solve for the ratio of uplift to erodibility (i.e. $[U/K] = S^n A^{-[m-1]}$) using slopes and drainage areas derived from DEMs. By calibrating this ratio to drainage basins where the assumption of homogeneous uplift appeared reasonable, they evaluated differences in erodibility for areas underlain by different lithologies. They found a four-fold variation in erodibility between areas underlain by erodible and resistant lithologies. They also evaluated spatial patterns of rock uplift rate after normalizing to account for these lithological effects.

In a similar approach, Kirby & Whipple (2001) analyzed downstream variations in θ to evaluate longitudinal gradients in rock uplift in the Siwalik Hills in central Nepal. They found that by assuming the river profiles were in steady state, erosion rates predicted by their calibrated stream power parameters implied rock uplift rates similar to those modeled by Hurtrez *et al.* (1999) from empirical relations between erosion rate and local relief. Consequently, in some cases area-slope characteristics of river profiles may provide insight into spatial patterns (and perhaps rates) of rock uplift across active geological structures.

Roe *et al.* (2002), however, showed that feedback between orographically variable precipitation and discharge-driven river incision implies that $\theta \neq m/n$ for steady-state landscapes with strong orographic precipitation regimes. Moreover, if the sediment flux through the reach is an important factor in controlling the rate of bedrock river incision (Sklar & Dietrich, 1998), as indicated by recent flume experiments (Sklar & Dietrich, 2001), then Eqs (2) and (3) become less pertinent to the field problem. Hence, as noted by Snyder *et al.* (2000), care needs to be taken in trying to infer m/n from observations of θ.

Influences of Vegetation. Studies in the past several decades have established that vegetation is a major influence on channel morphology and patterns at scales from individual channel units (e.g. pools and bars) to entire valley bottoms. Both live and dead vegetation influence channel form and processes. Live trees and grasses can contribute substantially to bank strength, and large woody debris (logs and logjams) can cause both local erosion and deposition (Fig. 2). Surveys of pool frequency in forest channels revealed that the majority of pools are forced by flow around wood, and the frequency of wood obstructions controls pool spacing (Beechie & Sibley, 1997; Buffington *et al.*, 2002, 2003; Montgomery *et al.*, 1995). The importance of bank vegetation has been demonstrated in a number of studies, and has long been recognized (e.g. Schumm & Lichty, 1965). In a series of field experiments, Smith (1976) demonstrated that plant roots can provide the dominant component of stream bank strength. Millar (2000) recently showed that the strength contributed by bank vegetation can be significant enough to influence the transition from a meandering to a braided channel pattern. Such an influence is apparent in evidence for a rapid change from meandering to braided river morphology coincident with the global plant die off 250 million years ago at the Permian/Triassic mass extinction event (Michaelsen, 2002; Ward *et al.*, 2000). The presence of grass or forest on river banks influences channel width, although whether channels widen or narrow depends on the type and geomorphic context of the channel (Davies-Colley, 1997; Stott, 1997; Trimble, 1997). Several studies have also investigated the role of vegetation on generating an anastomosing channel form by creating and maintaining local flow diversions that split a channel into a network of multiple channels (Collins *et al.*, 2002; Harwood & Brown, 1993; Tooth & Nanson, 1999). Such studies have broadened the range of scales over which vegetation is recognized as a primary influence on channel form.

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Fig. 2. Stable logjam acting as bank revetment on the Queets River, Washington.

Fig. 3. Log-filled reach of a small Pacific Northwest stream.

In recent decades, the distribution of in-stream wood and the accumulation of logs into log jams has been studied extensively in Europe and North America (Abbe & Montgomery, 1996; Downs & Simon, 2001; Gregory *et al.*, 1985, 1993; Gregory & Davis, 1992; Gurnell & Sweet, 1998; Piégay, 1993; Piégay & Marston, 1998; Robison & Beschta, 1990). In addition, researchers have established that large wood affects many channel processes (Fig. 3). In particular, recent studies have shown that wood influences channel roughness (Buffington & Montgomery, 1999a, b; MacFarlane & Wohl, in press; Manga & Kirchner, 2000; Shields & Gippel, 1995), bed-surface grain size (Buffington & Montgomery, 1999a, b; Lisle, 1995), pool formation (Abbe & Montgomery, 1996; Keller & Tally, 1979; Lisle, 1995; Montgomery *et al.*, 1995), channel-reach morphology (Keller & Swanson, 1979; Lisle, 1986; Montgomery *et al.*, 1996; Montgomery & Buffington, 1997; Nakamura & Swanson, 1993; Piégay & Gurnell, 1997), and the formation of valley-bottom landforms (Abbe & Montgomery, 1996; Collins *et al.*, 2002; Gurnell *et al.*, 2001). Some studies have shown that many of the geomorphic effects of wood in rivers arise from the influence of large stable wood as obstructions to flow and sediment transport (Abbe & Montgomery, 1996; Keller & Tally, 1979; Nakamura & Swanson, 1993). A number of workers have noted how the organization of wood and its effects on channels vary with position in the channel network (Abbe & Montgomery, 1996; Gurnell *et al.*, 2001; Keller & Swanson, 1979; Swanson *et al.*, 1982; Wallerstein *et al.*, 1997). Although many of the effects of wood occur at the scale of individual channel units (Bisson *et al.*, 1982), the integrated affects of these changes can alter channel properties at larger spatial scales of channel reaches and entire valley bottoms.

Disturbance Regimes. The characteristics and dynamics of stream habitat are recognized as providing a “geomorphic template” upon which aquatic ecosystems develop (Southwood, 1977). Disturbance regimes set by spatial and temporal variability in geomorphic processes capable of disrupting ecological systems or processes are viewed as a primary geomorphological control on stream ecosystems (Swanson *et al.*, 1988). The frequency, magnitude, and intensity of effects associated with a geomorphic process define its disturbance regime, and areas characterized by a similar disturbance regime define distinct process domains (Montgomery, 1999).

Episodicity is a fundamental characteristic of most geomorphic processes. Landslides and floods do not happen every day. The periodic nature of geomorphic phenomena has motivated ongoing examination of what controls the spatial and temporal scales over which steady-state assumptions may or may not apply. At the broad spatial and temporal scales of mountain range evolution, steady-state can be defined by constant exhumation rates over millions of years (Brandon *et al.*, 1998; Willett & Brandon, 2002). At this scale, individual storms may trigger catastrophic pulses of erosion but the long-term erosion rate is set by the rock uplift rate. The rise of mountains or periods of glaciation can act as disturbances over evolutionary time scales, but a single landslide can prove catastrophic to a local population of stream dwelling organisms. Although sediment transport in mountain streams is fundamentally episodic (Bunte & MacDonald, 1995), the time scales over which the integrated effects of discrete events can be meaningfully averaged depends on the nature of the analysis and the problem to be addressed. Incorporation of geomorphic processes into disturbance ecology, is reshaping understanding of aquatic and riparian ecosystem dynamics (e.g. Fausch *et al.*, 2002), in particular the role of periodic disturbances on the morphology and variability of mountain channel systems (e.g. Benda *et al.*, 1998).

Different kinds of organisms occupy different parts of a river system in part due to variations in the physical habitat template. Habitat characteristics and variability are influenced by the type, intensity, and frequency of disturbances. River systems exhibit both local and systematic downstream variability, as well as regional differences due to factors such as the geomorphic importance of small hydrologic events in a wet climate and the contrasting importance of rare events in arid climates. In mountain drainage basins, for example, headwater channels in confined valleys tend to be prone to high intensity, low frequency disturbances such as landslides and debris flows. After being scoured by debris flows such channels exhibit a temporal succession of habitat characteristics as material falls into the channel and gradually accumulates until the next scouring event. In contrast, lower-gradient alluvial channels in unconfined valley bottoms are frequently disturbed by lower intensity disturbances and by channel migration and avulsion. The style of disturbances in these different environments leads to differences in community structure and composition. Moreover, disturbance processes that may adversely impact local populations may be essential for creating and maintaining high-quality habitat in disturbance-prone environments. Consequently, understanding aquatic and riparian ecology may depend, in large part, on the integration of spatial and temporal disturbance processes and on their relation to the life history and distribution of particular organisms.

Post-Glacial Changes

Studies of post-glacial changes in river systems have improved understanding of regional river systems. In particular, legacies of Pleistocene glaciation on modern rivers have come to be appreciated. The oscillation between glacial and interglacial climates can result in sustained high sediment yields from rivers that never reach a steady state (Church & Ryder, 1972). Church & Slaymaker (1989), for example, showed how reworking of Pleistocene sediments still dominates the sediment budget for glaciated drainage basins in British Columbia.

Much of the progress in understanding post-glacial changes in riverine landscapes has been closely tied to advances in geochronology. Prior to the 1980s, late Quaternary geochronology was largely based on radiocarbon-dating or on relative dating using soils, stratigraphic position, rock weathering, or archeological context. Since the 1980s, numerical dating using cosmogenic isotopes, thermoluminescence, fission track, amino acid racemization, electron spin resonance and other techniques has been much more widely applied in Quaternary studies. These techniques have been especially useful in establishing chronologies for erosional or depositional episodes not directly associated with preservation of fossils. Many Holocene glacial chronologies for mountain ranges in the western U.S., for example, were originally based on radiocarbon ages from interbedded lake or marsh deposits. The use of cosmogenic isotopes to date glacial moraines directly has potential for improving the temporal resolution of glacial chronologies (e.g. Phillips *et al.*, 1990).

Coupled with advances in geochronology has been an increasingly quantifiable understanding of the episodicity of geomorphic change. The Pleistocene-Holocene transition was marked by enormous outburst floods from meltwaters ponded along the glacial margins. In the Channeled Scabland and northern margins of the Great Plains, these floods created landscapes that have been little modified by subsequent geomorphic processes (Baker & Nummedal, 1987; Kehew, 1993; Lord & Kehew, 1987). And, in regions as geologically and climatically diverse as the Appalachians and the Colorado Front Range, the Holocene was characterized at timescales of centuries to millennia by episodic geomorphic change driven by climatic variability. The relative importance of different styles of post-glacial change varied regionally across the United States, and these changes have left an imprint on modern river systems.

Puget Sound Rivers. The Puget Lobe of the Cordilleran Ice Sheet overran the Puget Sound about 17,000–16,000 cal yr B.P. (Porter & Swanson, 1998). As the Puget Lobe retreated northward at a rate of several hundred meters per year (Porter & Swanson, 1998), the melting ice exposed deeply incised valleys carved by sub-glacial streams. As river networks were re-established through a shifting network of spillways (Booth, 1994), some rivers came to occupy overdeepened subglacial meltwater troughs, whereas others were carved into the upland formed by the advance outwash. The modern character of Puget Sound rivers retains a legacy of these glacial origins (Booth *et al.*, 2003). Rivers flowing through sub-glacial meltwater troughs have aggraded during the Holocene, and were characterized historically by meandering channels (Fig. 4), some of which flowed through extensive valley bottom wetlands (Collins & Montgomery, 2001). In contrast, channels incised into advance outwash had few valley bottom wetlands and were characterized by an extensive network of anastomosing sloughs and side-channels (Collins & Montgomery, 2001). The distinction between these two contrasting styles of post-glacial history controlled the type and relative abundance of salmonid habitat in Puget Sound rivers at the time of Euro-American colonization.

Extensive post-glacial changes have also reshaped Puget Sound rivers. Post-glacial sea-level rise and isostatic rebound of up to 200 m in the North Sound have altered the extent of Holocene river valleys (Dethier *et al.*, 1995). Incision of rivers through the Holocene has altered the expanse of riverine valley bottoms (Beechie *et al.*, 2001). Immense mid-Holocene lahars from Glacier Peak created the extensive delta of the Skagit River (Dragovich *et al.*, 2000). In the 1980 eruption of Mount St. Helens, extensive lahars inundated the valley of the Toutle River, creating a broad valley flat (Fig. 5). Post-glacial establishment of forests further influenced Puget Sound rivers until historic clearing of snags and forest cover depleted in-stream wood and transformed the morphology of many Puget Sound rivers from complexes of anastomosing channels into relatively simple, single-thread meandering channels (Collins *et al.*, 2002). The post-glacial legacy has been one of extensive changes in Puget Sound rivers.

Mississippi River Drainage Basin. The retreat of the Laurentide ice sheet sent enormous volumes of meltwater flowing down the Mississippi River drainage network until the retreating ice sheet exposed the St. Lawrence and Hudson drainages. Recently-derived records of these meltwater floods come from $\delta^{18}O$ content in foraminifera (Joyce *et al.*, 1993) and grain-size variations of siliciclastic mud (Brown & Kennett, 1998) in the Gulf of Mexico, as well as from geomorphic evidence of channel cutting (Kehew & Lord, 1987; Knox, 1996) and alluvial fan deposition (Porter & Guccione, 1994). These records suggest that the ice sheet began to melt circa 14,000 ^{14}C yr B.P., with a meltwater megaflood from 12,600 to 12,000 ^{14}C yr B.P. (Brown & Kennett, 1998). Between 11,000 and 9,500 ^{14}C yr B.P., a rapid decrease

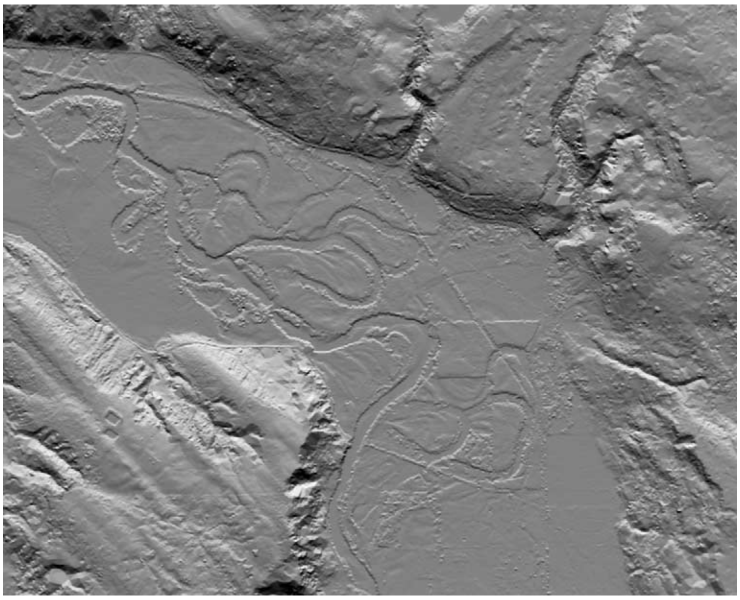

Fig. 4. Shaded relief map of LIDAR derived topography along the Snoqualmie River, Washington.

in discharge rate of the Mississippi drainage occurred as meltwater was directed eastward through the Hudson and St. Lawrence rivers (Broecker *et al*., 1989; Teller, 1990).

The changes in water and sediment yield associated with the latest Pleistocene glaciation caused large changes in the Mississippi River. The river initially incised in response to lowered baselevel during the height of the Wisconsinan glaciation (Schumm & Brakenridge, 1987). As the ice retreated, rapid drainage development occurred in the newly exposed land at the northern margin of the drainage basin (Anderson, 1988), and the central and lower portions of the river experienced aggradation and enhanced lateral movement. The channel in the lower basin began to change from braided to meandering ca. 8,800 ^{14}C yr B.P. (Baker, 1983).

The Holocene sedimentary record of the upper Mississippi River basin indicates fluctuations of ±30% of contemporary bankfull discharge, despite only modest changes in mean annual temperature and mean annual precipitation (Knox, 1993). During periods of larger floods (6,000–5,000 ^{14}C yr B.P., 3,300–2,000 ^{14}C yr B.P., A.D. 1450–1200), relatively rapid channel migration reworked or removed substantial amounts of valley-bottom alluvium (Knox, 1985). During periods of smaller floods (8,000–6,500 ^{14}C yr B.P., 5,000–3,300 ^{14}C yr B.P., and 2,000

Fig. 5. Lateral blast zone and lahar filled valley along the Toutle River at Mount St. Helens.

^{14}C yr B.P. to A.D. 1450), relatively slow lateral channel migration occurred and the channel and floodplain remained relatively stable (Knox, 1985). Overbank sedimentation on floodplains accelerated with the advent of agriculture in the region after A.D. 1820 (Knox, 1987).

The Holocene evolution of the lower Mississippi Valley has been a response to the effects of relative sea-level rise and variations in discharge and sediment delivery as driven by climate (Autin *et al.*, 1991). Individual rivers have alternately incised, aggraded, and changed their plan-view form during the Holocene (Autin, 1993), but regional stratigraphic records are not yet sufficient to determine whether these changes were broadly synchronous. Lobes of the Mississippi River delta, each approximately 30,000 km^2 and averaging 35 km thick, suggest that the delta's primary depositional site changes on average every 1,500 years (Coleman, 1988).

Rivers of the Colorado Front Range. The eastern edge of the Colorado Rocky Mountains from the Wyoming border south to the Arkansas River drainage constitutes the Colorado Front Range. The Front Range is drained by the channels of the South Platte River, which begin as mountain rivers confined within narrow bedrock canyons, and continue beyond the mountain-front as piedmont cobble- and gravel-bed rivers before becoming sand-bed channels farther east on the Great Plains. The post-glacial history of rivers in this region represents that of many mountain ranges in the Intermountain West in that the riverine landscape reflects erosional and depositional episodes over varyious timescales.

The piedmont along the eastern base of the Colorado Front Range has four pediment surfaces, the oldest of which is Pliocene in age, and a younger set of five strath or fill terraces of late Pleistocene and Holocene age (Morrison, 1987). The chronology for these surfaces was largely established during the 1960s using radiocarbon and relative geochronologic methods (Scott, 1960, 1963). The larger episodes of incision have been hypothesized to represent climatic changes, and the younger surfaces may reflect late Quaternary glaciation (Morrison, 1987). Front Range glacial chronologies based on radiocarbon and relative dating methods suggest between two and four glacial episodes during the Holocene (Benedict, 1973; Birkeland *et al.*, 1971; Burke & Birkeland, 1983; Richmond, 1960). More recent dating of Pleistocene glaciations indicate that Bull Lake moraines along upper Boulder Creek, one of the drainages in the Front Range, have minimum average ^{10}Be and ^{26}Al ages of 101,000 ± 21,000 and 122,000 ± 26,000 yr. Pinedale moraines along Boulder Creek have average model ages of 16,900 ± 3,500 yr and 17,500 ± 3,600 yr (Dethier *et al.*, 2000). Fill terraces downstream from the moraines along Boulder Canyon represent Bull Lake, Pinedale, and Holocene surfaces (Schildgen & Dethier, 2000). Limited cosmogenic and radiocarbon dating and soil development suggest that these terraces correlate with the terraces on the piedmont. Since ~600,000 yr ago, net incision rates on the High Plains near Boulder Creek have been $\sim 0.04\,\mathrm{mm\,yr^{-1}}$, whereas rates in Boulder Canyon have averaged $\sim 0.15\,\mathrm{mm\,yr^{-1}}$ since about 130,000 yr ago, suggesting that downcutting rates along the canyon have increased since early Pleistocene time (Dethier *et al.*, 2000).

In addition to erosional and depositional episodes driven by glacial and climatic change at timescales of thousands of years, rivers in the Colorado Front Range have undergone episodic change at timescales of hundreds of years as a result of hillslope instability and floods driven by precipitation and forest fires. Intense convective precipitation associated with summer thunderstorms can trigger slope mass movements and valley-bottom floods such as occurred in the Big Thompson River drainage during July 1976 (McCain *et al.*, 1979; Shroba *et al.*, 1979). Only these extreme floods, which recur at intervals of about 300–500 yr (Wohl, 2001), generate enough shear stress to overcome the high boundary resistance of the Front Range rivers. Such rare floods are thus very important in shaping valley and channel geometry.

Sierra Nevada Rivers. Once a matter considered decided, the topographic evolution of the Sierra Nevada is again controversial. According to classic studies of uplift of the Sierra Nevada, the range rose in post-Miocene or Pliocene time (Axelrod, 1957; Huber, 1981). Similarly, Wakabayashi & Sawyer (2001) argued that westward tilting, stream incision, and east-down normal and dextral faulting along the eastern escarpment of the range began ca. 5 myr ago. During these five million years, alpine glaciers repeatedly advanced and retreated, streams incised up to 1 km, and alluvial fan complexes developed along the mountain front. However, other recent studies, using new geochemical techniques, support the interpretation of little surface uplift in the Sierra Nevada since the early Tertiary (Chamberlain & Poage, 2000; House *et al.*, 1998).

Unlike the uplift history of the range, paleoclimate records of the Sierra Nevada have become less controversial in the past several decades. Pollen records indicate maximum glacial cooling of approximately 7–8 °C and, although precipitation inferences are less reliable, up to 2 m more annual precipitation during the glacial maximum (Adam & West, 1983). Radiocarbon and surface-exposure ages record multiple late Wisconsin advances in the Sierra Nevada (Osborn & Bevis, 2001), and glacial rock flour beneath Owens Lake suggests at least seven glacial advances between 84,000 and 15,000 yr ago (Bischoff & Cummins, 2001). Cosmogenic isotope ages from moraines in the eastern Sierra Nevada suggest that the transition from interglacial to full glacial conditions was rapid, with earlier glacial advances (ca. 200,000, 145,000, 115,000 yr ago) more extensive than later advances (ca. 65,000, 24,000, 21,000 yr ago) (Phillips *et al.*, 1990).

Pollen records from the Sierra Nevada indicate a drier climate 11,000–7,000 yr ago, a slight increase in precipitation 7,000–3,000 yr ago, and establishment of the present cool-moist climate after 3,000 yr ago (Anderson & Smith, 1994; Davis *et al.*, 1985). These climatic fluctuations have been associated with glacial advances during the latest Pleistocene, during an episode ca. 4,000–3,000 yr ago, and during the past several hundred years (Anderson & Smith, 1994). Sierra Nevada tree-ring records indicate that climate has remained relatively stable during the late Holocene (LaMarche, 1973) and that late Holocene hydrologic fluctuations are largely synchronous across the western United States (Earle, 1993).

An initial study if paleosalinity records from San Francisco Bay indicated no overall trends in the discharge of Sierra Nevada rivers during the past 2,700 yr (Ingram *et al*., 1996). However, average nonglacial erosion rates in the mountainous granitic terrain of the Sierra Nevada have varied by 2.5-fold during the Holocene (Riebe *et al*., 2001). Spatial variability in erosion rates across the Sierra Nevada has been attributed to proximity to fault scarps and river canyons. Erosion rates and hillslope gradients are strongly correlated at sites close to scarps and canyons. These sites appear to have accelerated local baselevel lowering and catchment erosion rates that are up to 15-fold higher than those of sites more distant from scarps and canyons, where erosion rates are much more uniform and less sensitive to average hillslope gradient (Riebe *et al*., 2000).

Rivers of the Sierra Nevada adjoin bouldery debris fans at the canyon mouths that commonly merge to form an alluvial apron. Relative fan size may reflect distribution of subsidence rates in the depositional basin (Whipple & Trayler, 1996), lithology and climate as these control both weathering rate and availability of unconsolidated material on canyon floors, and intense thunderstorm precipitation that generates sediment transport through flash floods and debris flows (Beaty, 1990; Bull, 1977). Alluvial fan deposition may be dominated by debris flows which determine both the structure of the channel network on the fan, and the long-term pattern of deposition on the fan suface (Whipple & Dunne, 1992). Alluvial fan deposition may also reflect sedimentation from outburst floods produced by failure of glacial moraines. Such fans are characterized by thick, unsorted, unstratified deposits that are a boulder-rich mix of clay to blocks deposited from noncohesive sediment gravity flows (Blair, 2001). Fans dominated by deposits from outburst floods lack the constituent levees, lobes and channel plugs, and alternating stacks of matrix-rich beds and washed gravel beds, present on fans dominated by debris flows (Blair, 2001).

Appalachian Rivers. The retreat of the Laramide ice sheet from the northeastern U.S. starting circa 16,000–15,000 ^{14}C yr B.P. was associated with rising baselevel for rivers draining to the Atlantic Ocean, glacial outburst floods from meltwater ponded along the margins of the ice sheet, and warming climate and associated changes in vegetation and weathering regime. Pollen and macrofossil records from the central Appalachians indicate an overall warming trend from 14,000 to ca. 7,500–6,000 yr ago (Kneller & Peteet, 1999; Webb *et al*., 1993). The northward expansion of boreal and temperate trees during this period produced many ephemeral forest communities. Today the southern Appalachians contain the most diverse tree flora of the eastern region (Davis, 1983). From a geomorphic perspective, probably the most important point is that the Appalachians and the eastern U.S. remained forested throughout the post-glacial period, so that rivers have responded to storms and floods rather than to changes in vegetative cover (Knox, 1983). Recent research in the Appalachians has tended to focus on (i) hillslope instability and the evolution of alluvial fans; (ii) the role of large floods in shaping contemporary river landscapes; and (iii) Cenozoic river incision and landscape evolution.

The late Pleistocene was a period of intense mechanical weathering and denudation in the Appalachian highlands (Clark & Ciolkosz, 1988; Mills & Delcourt, 1991); late Pleistocene slope denudation rates were an order of magnitude higher than Holocene rates (Braun, 1989; Saunders & Young, 1983). Sediment generated from creep and solifluction was stored in mountain hollows and episodically delivered to the valley floor by debris flows that on average recurred about every 2,500 yr (Eaton & McGeehin, 1997). Holocene warming terminated periglacial slope processes and reduced the rate of mechanical weathering. The reduction in sediment supply initiated stream incision through debris fans of late Pleistocene age, which resulted in lower, fans of Holocene age (Eaton, 1999). Landforms such as block fields and boulder streams that are relicts from late Pleistocene colder climates are now being modified by Holocene processes (Braun, 1989; Delcourt & Delcourt, 1988; Gardner *et al*., 1991).

Jacobson *et al*. (1989) emphasized two scales of temporal variation that influence hillslope instability and large floods in the central Appalachians: Quaternary climatic changes, and a higher frequency variation of rare events during the Holocene. The rare events arise from interactions between tropical storm paths and topographic barriers in the Appalachians. These interactions produce intense rainfall that may trigger hillslope mass movements and flooding. Topographically influenced flow concentration is the most important factor in determining relative slope stability throughout the region, but climate, lithology, geologic history and structure, and land use all exert an important influence (Mills *et al*., 1987). Comparison of the minimum precipitation threshold necessary to trigger debris flows in various regions of the U.S. indicates that longer and more intense rainfall is necessary in the Blue Ridge than elsewhere in the country (Wieczorek *et al*., 2000).

Hillslope instability in the Appalachians creates the alluvial fans that are the most prominent Cenozoic deposits in the region (Mills, 2000b). The instability also supplies debris that influences the evolution of channels and bottomlands (Miller, 1990; Mills *et al*., 1987). The Appalachians contain freely meandering streams, ingrown meandering streams confined within asymmetrical bedrock walls, and straight rivers within symmetrical bedrock walls that have little or no net valley-floor alluviation (Brakenridge, 1987). Although Appalachian valley bottoms do not preserve alluvial cut and fill cycles like those common throughout the southwestern U.S., Appalachian rivers do appear to have episodically enhanced rates of lateral channel migration, cutbank erosion and convex bend sedimentation that produce fill terraces (Brakenridge, 1987). The causes of such periodic enhanced erosion remain unclear, but certainly the occurrence of large floods plays a role. The geomorphic role of a flood varies in relation to drainage size. During a widespread flood in 1985, small, steep drainages scoured extensively; drainages of 1–65 km^2 had mixed erosion and deposition with continuous reworking of the valley floor; and drainages larger than 100 km^2 had only localized, discontinuous reworking (Miller, 1990). On a reach scale, the location and severity of flood impacts reflect longitudinal variations in valley width and channel orientation more than average width (Miller, 1995; Miller & Parkinson, 1993). And during moderate floods, basin geologic characteristics modify the severity of flooding, whereas the discharge of extreme floods is more closely controlled by precipitation characteristics resulting from storm motion and topographic features (Jacobson *et al.*, 1989; Smith *et al.*, 1996).

Research into Cenozoic river incision and landscape evolution in the Appalachians began more than a century ago (Davis, 1889). John Hack (1960) developed the concept of dynamic equilibrium in landscape evolution to explain his observations in the Valley and Ridge of Virginia. From Davis onward, geomorphologists have proposed dynamic incision of rivers into an asymmetrical mountain range, superposition from now-eroded overlying rocks, and superposition of river cutting through asymmetrical folds and thrust plates to explain the riverine landscapes in the Appalachians. Episodes of increased sedimentation in Mesozoic and Cenozoic marine basins (Poag & Sevon, 1989) may reflect periods of tectonic and/or climatic change, or divide migration and capture (Harbor, 1996). Investigations of individual rivers, including Virginia's New River and those to the north, have provided evidence of divide migration (Bartholomew & Mills, 1991; Hack, 1973).

Longitudinal profiles of most Appalachian rivers include distinct convexities where the streams traverse the Fall Zone, in which resistant rocks of the Piedmont bend downward beneath erodible rocks of the Coastal Plain (Pazzaglia & Gardner, 1994). Individual river incision rates include 0.027 mm yr^{-1} for the New River, Virginia (Granger *et al.*, 1997), 0.056–0.063 mm yr^{-1} for the Cheat River, West Virginia (Springer *et al.*, 1997), and 0.006–0.010 mm yr^{-1} for the Susquehanna River, Pennsylvania (Pazzaglia *et al.*, 1998). The rate of river incision may have increased during the late Cenozoic (Mills, 2000a). The incision history and terrace record of the Susquehanna River are the best-studied in the Appalachians.

The longitudinal profile and Miocene-Pleistocene age terraces of the Susquehanna River suggest complex interactions among relative baselevel, long-term flexural isostatic processes, climate, and river grade. Pazzaglia & Gardner (1993) proposed that the Susquehanna attained and maintained a characteristic graded longitudinal profile, such that bedrock straths were continually cut during periods of relative baselevel stability, with a change in climate or baselevel causing river incision and the formation of strath terraces. The Susquehanna strath terraces converge at the river mouth, diverge through the Piedmont, and reconverge to the north. This terrace profile deformation records progressive and cumulative flexural upwarping of the Atlantic margin (Pazzaglia & Gardner, 1993).

Anthropogenically Induced Changes

Despite the substantial post-glacial changes that have occurred in rivers throughout the United States, changes associated with human activities during the past two centuries have been so widespread and intense that some river systems have been more dramatically altered during this period than during earlier Quaternary climate changes. Channelization and levees have contained floodwaters and altered channel form; flow regulation has reduced peak flows and increased base flows, completely removed flow from a river channel, or driven a complete change in channel form; mining has induced massive increases in sediment transport and channel instability; and a variety of human activities have so loaded rivers with sediments, toxic contaminants, and excess nutrients that aquatic and riparian communities are impoverished in species diversity (NRC, 1992). Research during the past two decades has increasingly focused on human impacts to rivers and river landscapes as investigators have recognized how pervasive and substantial such impacts may be, even in apparently little-altered rivers (Wohl, 2001).

Snagging, Levees, and Channelization. Aboriginal forests across much of the United States were cleared throughout the 19th and 20th centuries (Fig. 6). Clearing of snags from the large rivers of the eastern and midwestern states was a matter of great commercial and military importance to the expanding country (Hill, 1957). In the 50 years after the first snagboat was built in 1829 to remove logs from the Mississippi and Ohio Rivers, more than 800,000 snags were pulled from the

Fig. 6. Forested buffer strip left after logging along a stream in the watershed of the Tolt River, Washington.

lower Mississippi alone. Over time, snagging extended to rivers throughout the Southeast and Midwest, and Pacific Northwest (Collins *et al.*, 2002; Sedell & Froggatt, 1984), where rivers were snagged and massive log rafts dismantled even before the valley bottom forests could be logged. The average diameter of cottonwood and sycamore snags from the Mississippi and Red Rivers exceeded 1.5 m (Sedell *et al.*, 1982; Triska, 1984); those pulled from rivers in western Washington were as large as 5.3 m in diameter (Collins & Montgomery, 2001). Records of snagging operations suggest wood loading in large Pacific Northwest rivers 100 times greater than now (Sedell & Froggatt, 1984), a difference similar to that estimated by comparing present-day wood loading in a protected reach of the lower Nisqually River to cleared reaches of the Stillaguamish and Snohomish rivers (Collins *et al.*, 2002).

After large rivers were cleaned of snags and jams in New England and the Pacific Northwest, tributary streams were catastrophically cleared through the ubiquitous practice of splash damming, in which a dam-break flood was induced to transport logs to the larger rivers from where they were then rafted to market. Splash damming was common throughout the Northwest (Sedell & Duvall, 1985), Intermountain West (Wohl, 2001), Midwest, and Northeast (Sedell *et al.*, 1982). These torrents scoured sediment and wood from streambeds and banks and reduced roughness and obstructions to flow, leaving many channels scoured down to bedrock.

As access to rivers increased, agriculture rapidly spread across adjacent floodplains. Within a matter of decades floodplain sloughs were typically ditched, valley-bottom wetlands drained, and side channels plugged (Fig. 7). Human occupation and use of floodplains made flooding a problem, so extensive networks of levees were built and many rivers were straightened, or channelized, in order to prevent flooding. An estimated 25,000 miles of levees enclose more than 30,000 square miles of floodplain in the United States (NRC, 1992). On large rivers, extensive construction of levees and dikes diminished floodplain storage of water during floods, thereby creating greater flow depths (higher stages of water) for the same discharge volume (e.g. Criss & Shock, 2001; Sparks, 1995). Flood control efforts in many instances exacerbated flooding hazards by promoting occupation of flood-prone areas (Fig. 8).

Cattle Grazing. Paleochannels and alluvial stratigraphy along rivers in the arid and semiarid regions of the Southwest and the intermountain West record numerous episodes of channel incision and aggradation throughout the Quaternary (Patton & Boison, 1986). The episodic instability of these dryland rivers became a focus of geomorphic research following an episode of widespread channel incision during the 1880s and 1890s (Graf, 1983). A lively debate has since persisted as to whether such channel change is driven by climatic variability, land use, or an intrinsic cycle of filling and entrenchment. Authors attributing channel change to land use have noted how intensive grazing results in lower vegetation density and higher runoff and sediment yields from uplands, as well as vegetation removal, trampling of banks, and associated reductions in bank stability along channels (Cooke & Reeves, 1976; Dodge, 1902; Leopold, 1921; Thornthwaite *et al.*, 1941). Before the mid-1940s, most studies attributed channel incision primarily to grazing. Consensus then shifted toward climatic variability as the main trigger for channel incision or aggradation (Graf, 1988). Various investigators concluded that (a) increasing precipitation created higher mean discharge, leading to greater erosive capacity; (b) decreasing precipitation led to a decline in stabilizing vegetation and allowed knickpoints to form and propagate upstream; or (c) changes in storm frequency led to periods of larger floods (channel incision) or smaller floods (aggradation) (Bryan, 1925; Hack, 1939; Leopold, 1951, 1976; Love, 1979). Since the early 1970s, episodic channel incision and aggradation have also been attributed to the inherent episodicity of channel processes along ephemeral rivers, where gradual aggradation by flows not competent to move sediment completely through a channel network eventually produces over-steepening, leading to formation of a headcut and channel incision (Patton & Schumm, 1975; Schumm & Hadley, 1957).


Fig. 7. Urbanized channel in the Puget Lowland, Washington.

The relative importance of climatic variability, land use, and intrinsic channel processes in regulating channel incision and aggradation remains a subject of contention and certainly varies among channels. However, there is consensus that intensive grazing within the riparian zone alters channel form and conditions. The net effect of grazing is that the channel

Fig. 8. Snoqualmie River, Washington, in flood during the winter of 1996.

becomes wider and shallower, has a finer substrate, less pool volume, less overhead cover from vegetation, undercut banks, warmer water and lower dissolved oxygen, more unstable banks, and less habitat diversity (Kauffman & Krueger, 1984; Magilligan & McDowell, 1997; Platts & Nelson, 1985; Trimble & Mendel, 1995). Changes in channel characteristics associated with grazing in the riparian zone include reduced shading and input of organic matter and reduced bank stability due to removal of riparian vegetation (Platts & Nelson, 1985; Trimble & Mendel, 1995), bank compaction, increased runoff, decreased infiltration, increased erosion due to trampling (Kauffman & Krueger, 1984; Magilligan & McDowell, 1997), and excess nutrient loads, lower dissolved oxygen, algal blooms and eutrophication from animal excrement (Behnke & Zarn, 1976; Trimble & Mendel, 1995). Grazing in the riparian zone is widespread on public lands in the western United States, where it may be the single greatest threat to the integrity of aquatic habitat (Behnke & Zarn, 1976).

Flow Regulation

Changes in the flow regime – the magnitude, frequency, and duration of flow – along a river may result from diverse human activities including dams, diversions, groundwater withdrawal, and urbanization. Many of these activities, and the concomitant channel changes, are now ubiquitous across the United States, although the specific impacts vary by region. In some regions of North America dams have had a greater effect on rivers and aquatic ecosystems than Quaternary climate changes.

The effects of flow regulation depend on the associated changes in flow regime. The effects of dams, for example, depend on the purpose to which the dam was built. Dams operated for flood control, hydroelectric power generation, or water storage commonly decrease peak flows, result in strong fluctuations over 24-hr period (hydroelectric) or shift in timing of seasonal peak (water storage), decrease downstream sediment supply, leading to channel bed armoring or erosion, and channel instability, and may change water temperature and chemistry (Graf, 1996; Hirsch *et al*., 1990; Williams & Wolman, 1984). Flow diversions may remove or add water to channel, and change the magnitude and timing of flow. Where water withdrawal is so pronounced that channel form (e.g. substrate grain-size, pool volume, width/depth ratio, flood conveyance), habitat, or recreational uses are impaired, efforts may be made to purchase or legislate some minimum volume of flow within the channel (instream flow), specified as an annual minimum or as minimum flows at various times during an annual hydrograph (Schleusener *et al*., 1962; Stromberg & Patten, 1990). Groundwater withdrawals can lower the local or regional water table to the degree that in extreme cases base flow to a river may be reduced or even cause a perennial river to become intermittent or ephemeral. Water withdrawals may also cause channel incision, particularly if the withdrawal causes regional compaction or subsidence (Kondolf, 1996). Changes in flow due to increased impermeable surface in a drainage basin increases the magnitude and rate of runoff for a given precipitation input (Morisawa & LaFlure, 1979; Urbonas & Benik, 1995; Wolman, 1967). This commonly increases the peak flow of small to moderate floods, and makes flood hydrographs more flashy. Combined with a decrease in sediment yield, such changes increase channel erosion and instability.

Pl. check reference "Behnke & Zarn (1976)" which is missing in reference list.

The continental U.S. has 75,000 dams that together are capable of storing a volume of water almost equaling one year's mean runoff (Graf, 1999). The greatest impacts to river flow from dams occur in the Great Plains, Rocky Mountains, and the arid Southwest, where storage is almost 4 times the mean annual runoff (Graf, 1999). In these regions and elsewhere, such as the Columbia River basin, dams have been identified as a major impact on native fish species (Ligon *et al*., 1995). Surface-water withdrawal for offstream uses such as irrigation is greatest in California, Idaho, Colorado, and the western Great Plains.

The Mississippi River presents an example of how Holocene patterns of water and sediment discharge have been altered by dams, diversions, and channel structures constructed during the 20th century. In the 208,000 km^2 of the upper Mississippi River basin, more than 13,000 km of levees have been built, and 65% of the original wetlands have been drained. By 1950, a system of 29 locks and dams had been built along the upper Mississippi River from St. Louis to Minneapolis, and a 2.7-m navigation channel had been dredged from St. Louis to Sioux City, Iowa (Watson & Biedenharn, 2000). Although the Mississippi River historically had the greatest water and sediment discharges in the U.S. (Meade *et al*., 1990), the storage of large volumes of sediment behind more than 8,000 dams (Graf, 1999) is causing various downstream impacts, including subsidence and erosion of the Mississippi delta and adjacent coastlines (Britsch & Dunbar, 1993).

Some of the most dramatic changes in the river landscape as a result of flow regulation have occurred in the western Great Plains. Rivers such as the South Platte, the North Platte, and the Arkansas, were broad, shallow, braided channels when people of European descent first described them in the 19th century (Eschner *et al*., 1983; Williams, 1978). Phrases such as “a mile wide and an inch deep” and “too thick to drink but too thin to plow” were used to describe these rivers which flowed high with late spring-early summer snowmelt, then shrank back to very low flows by autumn. The channels had sparse riparian vegetation and warm, turbid water. Beginning at the end of the 19th century, reductions in the snowmelt flood peak, increased late summer base flow, and higher regional water tables resulting from flow regulation and extensive agricultural irrigation, facilitated the establishment of riparian trees. Within a few decades, braided channels that had been 450 m wide became 150-m-wide sinuous channels with densely vegetated islands and banks (Nadler & Schumm, 1981).

The flow regime of most U.S. rivers has been modified to some degree by human actions. Growing recognition that aquatic ecosystems can be finely adjusted to the hydrologic regime, and the channel features and habitats that it creates or sustains, is focusing attention on how to set flow regimes in regulated or managed channels so as to maintain channel form and ecological functions. Approaches to setting instream flows in managed or regulated rivers are generally based on habitat preferences that can be characterized in terms of flow depth or velocity. But as higher flows are generally required to form and maintain habitat, a major concern with such approaches for determining minimum flows in regulated rivers is that they specify flows needed to maintain the use of habitats but not the habitats themselves (Whiting, 2002). Although there is no simple way to determine the flows needed to maintain a channel, it has become clear that the closer the annual hydrograph is to the natural flow regime, the more likely it is that flow-habitat interactions will be ecologically effective and sustainable (Poff *et al*., 1997; Whiting, 2002).

Mining. Mining may occur within the river corridor, as when placer metals disseminated through valley-bottom sediments are removed, or alluvial sand and gravel deposits are mined for construction aggregate. Mining may occur elsewhere within a watershed if lode metals disseminated through bedrock outcrops are mined, or if fossil fuels such as coal and oil are removed.

Impacts of mining on rivers generally vary with the type of mine. Placer mining decreases bed and bank stability, increases downstream sediment transport, and reduces water quality. Associated toxins such as mercury, increased sediment transport and channel instability can each stress or destroy aquatic and riparian organisms (Hilmes & Wohl, 1995; James, 1991, 1994, 1999; Van Haveren, 1991; Van Nieuwenhuyse & LaPerriere, 1986). Aggregate mining decreases bed and bank stability, and depressions created by mining may initiate headcut erosion, trap sediment and create sediment depletion downstream, or divert flow and cause lateral channel movement, which increases downstream sediment transport and thus reduces water quality and alters aquatic habitat (Bull & Scott, 1974; Chang, 1987; Kondolf, 1994, 1997; Lagasse *et al*., 1980; Norman *et al*., 1998). Lode mining may increase sediment yield to channels from tailings and from slope instability associated with deforestation, and may cause acid-mine drainage (Starnes & Gasper, 1995; Stiller, 2000; Wohl, 2001). Finally, strip-mining may completely alter topography and water and sediment yields, or even obliterate streams (as, for example, in mountain-top removal in West Virginia). In addition, contaminants in water used to mine or process fuels may completely alter local river flow regime (e.g. coal-bed methane mining in the intermountain West) (Starnes & Gasper, 1995).

Placer mining in the continental U.S. occurred mainly in the intermountain West during the 1850s–1950s. California’s Sierra Nevada, Colorado’s Front Range, western Montana, western Nevada, and the Black Hills of South Dakota were among the regions with the most intense placer mining (Fig. 9). The massive amounts of sediment mobilized by mining activities, as well as associated changes in flow regime and introduction of toxic contaminants, continue to affect these rivers (Alpers & Hunerlach, 2000; Hilmes & Wohl, 1995; James, 1991, 1999; Stiller, 2000).

Aggregate mining for sand and gravel is widespread in the U.S. (Tepordei, 1987). In 1990, approximately 4,200 companies mined 830 billion kg of sand and gravel from 5,700 operations along rivers and floodplains (Meador & Layher, 1998). Nearly all of this material is used in construction, usually within 50–80 km of the mine. In-channel mining occurs as: (a) dry-pit mining in which a pit is excavated below the thalweg of a dry ephemeral stream; (b) wet-pit mining in which the pit extends below the water table; (c) bar skimming in which all the sediment in a gravel bar above an imaginary line sloping upwards from the summer water’s edge is removed; or (d) safe yield in which extraction is limited to

Fig. 9. Hydraulic mining at the Malakoff diggings of the North Bloomfield Gravel Mining Company in the Sierra Nevada (Plate A from Whitney, 1880).

the removal of annual aggradation (Kondolf, 1994, 1997; Sandecki, 1989). Floodplain and terrace pit mining occur in fluvial deposits beyond the active channel. Instream mining commonly causes channel incision that may propagate up- and downstream from the mine, undermining structures such as bridges. Incision may also induce channel instability that changes substrate grain-size distribution and bedform configuration, and causes downstream siltation and reduced water quality. And incision may lower valley-bottom water tables and destroy riparian environments and hyporheic exchanges (Kondolf, 1997; Norman *et al.*, 1998; Sandecki, 1989).

Lode mining has focused on industrial metals such as iron in the upper Great Lakes, and precious metals such as silver and gold in the southern Appalachians and the West. Most active contemporary lode mining occurs in the western U.S., and the majority of this region's toxic waste sites are associated with historical or contemporary mining. Impacts to rivers adjacent to lode mining derive primarily from increased sediment yields and introduced toxic contaminants (Rampe & Runnells, 1989; Stiller, 2000; Stoughton & Marcus, 2000).

Coal has been mined extensively in the Appalachians, the upper and central Midwest and Great Lakes region, the intermountain West and the upper Great Plains. Oil in the continental U.S. has come primarily from the southern Great Plains and Gulf Coast, the Mississippi Valley, and parts of the northwestern Great Plains. Natural gas came from the Appalachians, the Mississippi Valley, and the southern and western Great Plains (National Geographic Society, 1998). Some of the most active contemporary mining for fossil fuels occurs in the western Great Plains (e.g. Wyoming), the South (e.g. Texas and Louisiana), and the Appalachians (e.g. West Virginia). The effects on rivers of such mining vary in association with the type of mining, but commonly involve increased sediment yields and the introduction of toxic contaminants.

Water Quality. Most rivers in the continental United States have contaminants or reduced water quality resulting from human activity. The Clean Water Act of 1972 set a "fishable" and "swimmable" goal for all waterways. Over the next twelve years the federal government contributed a third of the $310 billion spent to clean up surface waters. Spending on water pollution remained high in subsequent decades; in 1992, for example, the Environmental Protection Agency spent $2.9 billion, largely in the form of grants to states for sewage treatment plants. By 1994, such programs had reduced sewage in American rivers by 90% relative to 1970 (Vileisis, 1999). However, the U.S. did not, and has not, come close to meeting the 1985 target date of the Clean Water Act for fishable and swimmable waters everywhere.

Pl. check reference "Vileisis (1999)" which is missing in reference list.

Pollutants in streams represent have various sources, and present differing potential hazards to humans and other organisms. Elevated sediment loads derived from runoff from agricultural and otherwise managed lands can change channel substrate and form and destroy aquatic and riparian habitat (Waters, 1995). Excess nutrients (N, P) from fertilizers or sewage systems (animal waste, laundry detergent) can lead to excessive algal growth and low levels of dissolved oxygen, or introduce carcinogenic by-products of treatment in chlorinated drinking water (Graffy *et al.*, 1996; Steingraber, 1997). Trace elements (such as As, Cd, Cr, Cu, Pb, Hg, Ni, Se, Zn) can be introduced by atmospheric deposition (volcanic emissions; combustion of municipal solid waste and fossil fuels in coal- and oil-fired power plants; releases from metal smelters, automobile emissions, biomass burning), point source releases to surface water (municipal sewage sludge, effluent from coal-fired power plants, releases from industrial uses, acid-mine drainage), or nonpoint source releases (natural rock weathering, agricultural runoff of manure and artificial fertilizers, wear of automobile parts, irrigation return flow). Many trace elements are adsorbed to fine sediments, taken up by invertebrates, and passed through the food web. Individual elements may bioaccumulate (accumulate within the body of an organism) and biomagnify (concentrate as they are passed between organisms). Trace elements are commonly teratogens (cause developmental changes and abnormalities), mutagens (cause chromosomal changes) and carcinogens (cause cancerous growths) (Rice, 1999). Organochlorine compounds (pesticides, PCBs) from agricultural and municipal application of herbicides and insecticides; waste from electricity-generating facilities, and many common items (e.g. photocopy toner) in wastewater effluents and atmospheric fallout from incinerators, or that leach from landfills can have a range of behavior (mobility, persistence, toxicity) that varies widely among individual compounds. The worst-case scenarios are compounds such as DDT, the toxic breakdown products (DDE) of which are still present 30 years after the last application in the U.S., and which act as endocrine-disrupters in many species (Colborn *et al.*, 1997; Nowell *et al.*, 1999; USGS, 1999). The behavior and health effects of volatile organic compounds (VOCs) introduced into streams from nonpoint sources (primarily urban land surfaces and urban air) are variable and largely unknown, but this large group of more than 60 compounds includes diverse substances and carcinogens such as benzene, the solvent terachloroethylene, toluene, and chloroform (Lopes & Bender, 1998; Rathbun, 1998).

Sediment remains the most important river pollutant in terms of number of stream miles degraded (Waters, 1995), and excess fermentable organic wastes from human and animal sewage create locally significant impacts on water quality. The most insidious contaminants come from industrial and agricultural activities. The U.S. Geological Survey's National Water Quality Assessment (NAWQA) program, begun in 1991, provides a comprehensive index of national water conditions. The first phase of the program included 59 study units throughout the continental U.S., Alaska and Hawaii. Standardized sampling of surface and ground waters within these study units assessed water chemistry, streambed sediments, invertebrate and fish tissue, and stream habitat. Sample testing included analyses for 9 trace elements, 33 organochlorine compounds and 106 pesticides, 5 nutrients, and 60 volatile organic compounds. Every one of the 59 study units sampled had some type of contaminant that locally exceeded either drinking-water standards, or standards for the protection of aquatic life. Even forested watersheds with little direct land use contain residues of such synthetic compounds as DDT or PCBs, which reached the watershed from atmospheric sources, although the use of DDT was banned in the U.S. in 1972, and the use and manufacture of PCBs was banned in the U.S. in 1979. The cumulative impact of these contaminants impairs aquatic and riparian ecosystem functioning by reducing the diversity and abundance of organisms in river landscapes.

Summary of Anthropogenic Changes. Other than rives once overrun by glacial ice, anthropogenic changes exceed those due to Quaternary climate changes in many river systems. Hooke (1999, 2000) estimated that throughout much of the U.S., humans now move more sediment (average 31,000 kg yr^{-1} per capita) than do rivers. The net result of a diverse array of human activities is to move rivers and riverine landscapes toward increasing homogeneity and reduced environmental quality. We straighten and deepen channels; confine floodwaters and stop the processes maintaining floodplains and riparian corridors; increase sediment movement and alter channel bedforms and planform; reduce peak flows and increase base flows; and poison water and sediment with adsorbed toxics. All of this moves diverse, stable, functional river ecosystems toward a condition in which river form and process are so altered that channels begin to resemble irrigation canals or drainage culverts.

Emerging Research Directions

Much remains to be learned about rivers in general, and about the status, behavior, and history of particular river systems. In particular, we see four emerging areas of research interest as providing significant new opportunities: river restoration and rehabilitation, biogeomorphology, resistant-boundary channels, and interactions among tectonics, climate, and erosion.

River Restoration and Rehabilitation. River restoration programs around the United States aim to improve the quality of rivers for use by both humans and wildlife. Rivers are dynamic systems in which specific attributes are continually created, altered, and destroyed. Consequently, river restoration means not only reestablishing certain prior conditions but also reestablishing the processes that create those conditions. In contrast, river rehabilitation aims to improve river conditions but does not necessarily seek reestablishment of natural conditions and dynamics. Given the extensive historic changes to rivers, and the resulting constraints, most projects billed as "river restoration" actually achieve only a form of river rehabilitation.

Techniques being used to rehabilitate rivers include setting levees back away from channel banks to allow the channel to migrate within a proscribed corridor on its historical floodplain. Delineation of channel migration zones and erosion hazard zones also are starting to be used in regulatory arenas to account for the potential for channel movement that could affect long-term capital projects or impact the assumptions or objectives underpinning forest-harvest planning. Streamside buffer zones have been widely applied to forestlands and are now being adopted in some urbanized settings. The central importance of the natural flow regime in stream ecology also is becoming recognized in stream restoration and rehabilitation programs and projects.

Biogeomorphology. Disturbances are generally considered to negatively impact aquatic ecosystems, but catastrophic disturbances such as floods and landslides also can locally create or enhance aquatic and riparian habitat (Everest &

Fig. 10. Salmon swimming through the Ballard Locks in Seattle.

Meehan, 1981; Friedman & Auble, 2000a, b; Reeves *et al*., 1995). Consequently, the net benefit or detriment to an aquatic population will depend on the type, frequency, and impact of a disturbance and its relative importance in creating essential characteristics for that species. To assess the relative role of disturbance processes on aquatic ecosystems habitat, one must consider the net effect of habitat destruction and creation. Recognizing that an organism evolved in a dynamic environment does not necessarily imply that disturbance processes are essential for maintaining that organism. Understanding the effects of disturbance processes on populations of organisms requires understanding the full distribution of events and their effects across space and through time. For most systems and organisms such understanding remains in its infancy (Fig. 10).

Resistant-Boundary Channels. Resistant-boundary channels are those formed on bedrock or on very coarse clasts, such as rivers in mountain regions formed on boulders. These types of channels are not adequately described by the conceptual and mathematical models commonly applied to lower gradient alluvial rivers (Tinkler & Wohl, 1998; Wohl, 2000). Resistant-boundary channels tend to have steep gradients and rough boundaries that resist fluvial erosion. These channels have highly turbulent flow; supply-limited sediment transport; highly stochastic sediment movement that is difficult to parameterize and model; abrupt downstream variation in channel geometry; and episodic channel change restricted to the relatively high magnitude, low frequency flows that are capable of exceeding erosional thresholds along bedrock rivers and mountain rivers (Baker, 1988; Tinkler & Wohl, 1998; Wohl, 1998).

Bedrock rivers have increasingly been the subject of research because knowledge of processes and rates of bedrock channel incision is vital to quantitative modeling of landscape evolution. River channels are the conduits by which weathering products are removed from a drainage basin. This rate of removal influences the efficiency of landscape change. For example, Burbank *et al.* (1996) describe a balance between channel incision and hillslope profiles in the Indus River drainage basin such that, as a bedrock channel incises, adjacent hillslopes become over-steepened, triggering mass movements. The resultant influx of sediment to the channel decreases the rate of channel incision until the sediment has been transported downstream, at which time a new period of incision occurs. The rate and manner of bedrock channel erosion thus partly control hillslope stability and evolution of the entire catchment area.

Many recent studies have focused on quantifying the variables controlling erosional processes and channel geometry along bedrock channels, as well as the resultant long-term incision rate (Howard, 1998; Roe *et al*., 2002; Seidl *et al*., 1994; Snyder *et al*., 2000; Stock & Montgomery, 1999; Whipple & Tucker, 1999). The findings begin to delineate the conditions under which specific channel incision regimes occur. Stock & Montgomery (1999), for example, have proposed that a weak dependence of incision rate on drainage area characterizes areas where abrupt baselevel fall produces channel incision primarily through knickpoint retreat. In contrast, rates of channel incision under conditions of stable baselevel depends strongly on drainage area. Recent studies also suggest that bedrock channel geometry responds consistently to the balance between hydraulic driving forces and substrate resisting forces, such that bedrock channel geometry may be predictable in some circumstances (Montgomery & Gran, 2001; Wohl & Merritt, 2001).

Investigations have increasingly focused on mountain rivers in recognition that these headwater stream segments produce a disproportionately large component of the sediment yield from a drainage basin (Milliman & Syvitski, 1992). However, adequate equations do not yet exist to describe hydraulics and sediment transport along mountain rivers. Steep gradients and large grain and form roughness promote non-logarithmic velocity profiles (Wiberg & Smith, 1991), localized critical and supercritical flow, and strongly three-dimensional flow in these rivers (Wohl, 2000). Recent work on the hydraulics of mountain rivers has attempted to: (a) predict flow resistance coefficients as a function of gradient, relative submergence, flow depth, particle size distribution, or other channel characteristics (Jarrett, 1990; Marcus *et al.*, 1992; Maxwell & Papanicolaou, 2001); (b) quantify the contribution of the components of grain and/or form roughness to total flow resistance (Bathurst, 1993; Curran & Wohl, in press; Prestegaard, 1983); and (c) characterize cross-stream velocity and vertical velocity distribution and the associated forces of lift or shear stress exerted on the channel boundaries (Furbish, 1993; Wohl & Thompson, 2000).

Students of mountain channels have barely begun to understand how sediment in transport variously shields and scours river beds (e.g. Sklar & Dietrich, 1998, 2001). Spatial and temporal variations in hydraulics and bed substrate, and limited sediment supply, render bedload entrainment and transport equally difficult to characterize in mountain streams. Research has focused on (a) the grain-size distribution of channel-bed sediments (Buffington & Montgomery, 1999a, b; Ferguson & Paola, 1997; Nikora *et al.*, 1998; Wohl *et al.*, 1996; Wolcott & Church, 1991); (b) sediment entrainment (Johnston *et al.*, 1998) and the occurrence of equal mobility vs. selective entrainment (Kuhnle, 1992; Montgomery *et al.*, 1999; Wathen *et al.*, 1995; Wilcock, 1993); and (c) the mechanics of bedload transport (Gomez & Troutman, 1997; Thompson *et al.*, 1999; Wilcock *et al.*, 1996).

Other areas of recent focus in mountain rivers include bedforms and channel morphology, and longitudinal profile development. Flow energy and sediment supply likely interact along a continuum of channel-bed gradient to produce predictable trends in channel bedforms and channel morphology (Chin, 1998; Montgomery & Buffington, 1997), but conceptual models describing these trends are presently limited by a lack of field-based data describing how channel morphology varies as a function of potential controls. Similarly, a quantitative understanding of the controls on spatial distribution and relative importance of different processes of incision along mountain rivers requires further detailed, extensive field measurements of hillslope and channel processes in mountainous regions (Wohl, 2000).

Tectonics, Climate, and Erosion. An exciting area of active research focuses on coupling and feedback among climate, erosion, and tectonic processes. Over the past decade geologists have recognized that the development and evolution of geologic structures can depend on spatial gradients in the climate forcing that drives erosion (Avouac & Burov, 1996; Hoffman & Grotzinger, 1993; Horton, 1999). Models that couple geodynamics and surface processes in evolving and steady-state orogens, and which predict the resulting metamorphic gradients exposed at the surface, reflect the influence of spatial variability in surface erosion (Willett, 1999; Willett *et al.*, 1993, 2001). Development of mountain ranges strongly influences patterns of precipitation (Barros & Lettenmaier, 1994) and therefore patterns of erosional intensity, which in turn governs the development and evolution of topography. Steady-state river long profiles, for example, are influenced by orographic controls on precipitation patterns (Roe *et al.*, 2002). The role of fluvial erosion in reducing mass accumulation in mountain ranges is perhaps best illustrated by the converse cases where lack of rainfall, and therefore limited erosion, allows accumulation of enough mass to engage the mechanical limit to crustal thickening (Pope & Willett, 1998) and result in development of high plateaus like the Altiplano and Tibet (Montgomery *et al.*, 2001) (Fig. 11). Climate, erosion, and tectonics are thus coupled through feedbacks. How this leads to positive feedback between erosion and tectonics provides a fruitful avenue for further inquiry.

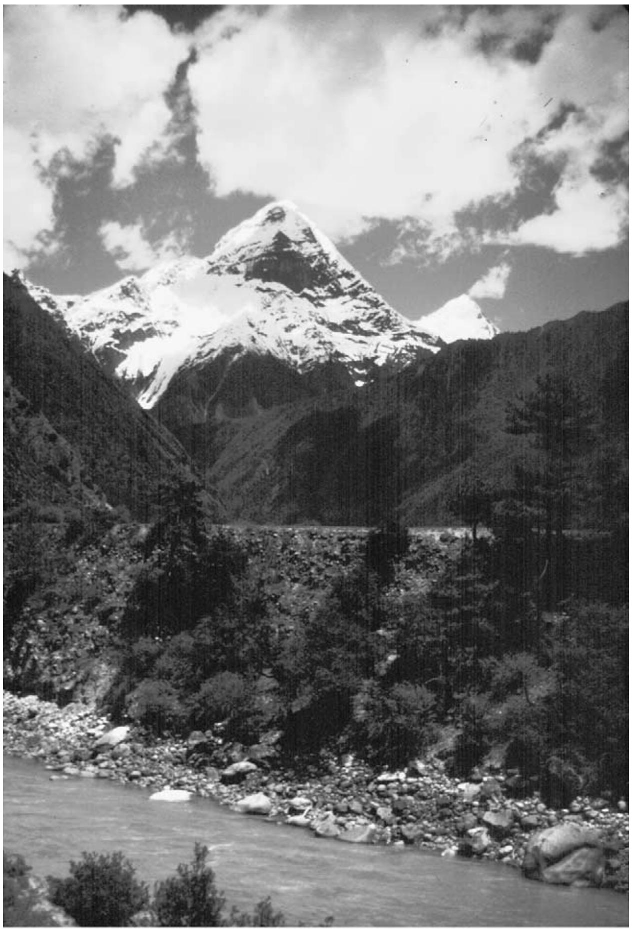

Fig. 11. High peak along a river near Namche Barwa in eastern Tibet.

Summary

The discipline of fluvial geomorphology has broadened as the science has developed over the past 40 years. Understanding of the fundamental aspects of rivers and the variability in fluvial processes and conditions due to regional differences in climate and geologic histories has advanced considerably. Connections to other disciplines are coming to the forefront of research, and offer exciting potential for further connections and for investigation of feedback between both system components and systems traditionally treated as separate domains. The investigation of resistant-boundary channels is an exciting and under-explored area of fluvial geomorphology that challenges us to rethink traditional assumptions based on alluvial channels. The dramatic influence of fluvial processes on aquatic and riparian ecology paves the way for collaboration with stream ecologists. Modern geobiology, encompassing the interactions of organisms and their environment, offers frontier opportunities to researchers willing to work at the interface between geomorphology and biology. Over vastly expanded temporal horizons the role of erosion as a geological process remains an area with exciting research potential. Finally, the pressing need for better integration of fluvial geomorphology (and fluvial geomorphologists) in river restoration and rehabilitation presents an important challenge to the research community.

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