

CHAPTER 5

THE CHANNEL BED – CONTAMINANT TRANSPORT AND STORAGE

5.1. INTRODUCTION

In the previous chapter we were concerned with trace metals occurring either as part of the dissolved load, or attached to suspended particles within the water column. Rarely are those constituents flushed completely out of the system during a single flood; rather they are episodically moved downstream through the processes of erosion, transport, and redeposition. Even dissolved constituents are likely to be sorbed onto particles or exchanged for other substances within alluvial deposits. The result of this episodic transport is that downstream patterns in concentration observed for suspended sediments is generally similar to that defined for sediments within the channel bed (although differences in the magnitude of contamination may occur as a result of variations in particle size and other factors) (Owens et al. 2001). While this statement may seem of little importance, it implies that bed sediments can be used to investigate river health as well as contaminant sources and transport dynamics in river basins. The ability to do so greatly simplifies many aspects of our analyses as bed materials are easier to collect and can be obtained at most times throughout the year, except, perhaps during periods of extreme flood. Moreover, bed sediment typically exhibits less variations in concentration through time, eliminating the need to collect samples during multiple, infrequent runoff events. Sediment-borne trace metals are not, however, distributed uniformly over the channel floor, but are partitioned into discrete depositional zones by a host of erosional and depositional processes. Thus, to effectively interpret and use geochemical data from the channel bed materials, we must understand how and why trace metals occur where they do.

In this chapter, we will examine the processes responsible for the concentration and dispersal of sediment-borne trace metals, and the factors that control their spatial distribution within the channel bed. We begin by examining the basic mechanics involved in sediment transport, including an analysis of open channel flow. We will then examine how these basic processes produce spatial variations in sediment-borne trace metal concentrations, both at the reach scale and at the scale of the entire watershed.

5.2. SEDIMENT TRANSPORT

5.2.1. Modes of Transport

Our understanding of sediment transport processes is hampered by the difficulties of obtaining measurements in natural channels. Rivers are characterized by ever changing discharge and velocity and often are bounded by erodible, inhomogeneous bed and bank materials. In addition, river water is usually opaque, thereby making visual examination of the channel bed impossible. In order to overcome these difficulties, investigators have supplemented field investigations with laboratory studies that allow the transport of sediment to be observed in flumes where slope and discharge can be changed, and other selected variables can be held constant. These laboratory studies have been instrumental in deciphering the mechanisms responsible for initiating particle motion, and for predicting the relationships between sediment transport rates and discharge. Nonetheless, our ability to predict the size and quantity of sediment (and thus, contaminants) that a river can carry under a given set of flow conditions remains less than perfect.

The movement of sediment in a channel varies with both time and space. It is a function of the size of the sediment that is being transported as well as the energy that is available to perform mechanical work. In the previous chapter, we examined the downstream flux of fine-grained particles transported within the water column as part of the suspended load. These suspended particles have little, if any, contact with the channel bed. They are held in suspension by short, but intense, upward deviations in the flow as a result of turbulent eddies which are generated along the channel margins. The more frequent and intense the turbulence, the more material, and the larger the grains, that can be held within the water column (Mount 1995). We generally think of the suspended load as consisting of only fine grained sediment. However, during extreme floods in steep, gravel bed-rivers, particles a few tens of centimeters in diameter have been observed to move in true suspension.

Suspended sediment typically moves at a rate that is slightly slower than that of the water, and may travel significant distances downstream before coming to rest. In contrast, *bedload*, which is transported close to the channel bottom by rolling, sliding, or bouncing (saltating), is unlikely to be moved great distances before it is deposited and stored, either temporarily or semi-permanently, within the channel. It is important to recognize that sediment which moves as suspended load during one flood event, may move as bedload during another, depending on the hydraulic conditions to which the particles are subjected.

The ability of rivers to transport sediment is described in terms of competence and capacity. *Competence* refers to the size of the largest particle that a river can carry under a given set of hydraulic conditions. *Capacity* is not concerned with the size of the material, but is defined as the maximum amount of sediment that a river can potentially transport. Note that capacity represents a theoretical maximum. As such, it is almost always larger than the actual sediment load being carried by the river.

It should come as no surprise that as discharge increases the amount of sediment being transported also will increase. The relationship, however, is much more complicated than one might think. For example, the correlation between the suspended load and discharge is generally quite poor because rivers can typically transport more fine-grained sediment than is supplied from the watershed; the load is controlled by supply rather than discharge. In contrast, the amount of material available for transport as bedload generally exceeds that which can be carried by the river. The correlation between discharge and bedload transport rates, then, is much better than for suspended load, but it is still less than perfect. In the case of bedload, the correlation is significantly influenced by complexities associated with particle entrainment (discussed below). Another important problem is that bedload transport is difficult to measure because the movement of particles along the channel bed is primarily associated with floods when flow depths and velocities are elevated. Working in these conditions can be difficult and dangerous. In addition, where bedload transport rates have been accurately measured along a heavily instrumented river, they are found to vary across the width of the channel bed and through time (Leopold and Emmett 1977; Hoey 1992; Carling et al. 1998). These variations are at least partly related to the movement of large-scale bedforms along the channel floor. The periodic movement of a bedform through a cross section hinders the use of hand held instruments to measure bedload transport rates because they can only sample a single location for a short period of time.

The difficulties of directly measuring bedload has led to the use of mathematical models to estimate a river's load for given a set of channel, sediment, and flow conditions (Meyer-Peter and Muller 1948; Einstein 1950; Bagnold 1980; Parker et al. 1982). The accuracy of the estimated transport rates can be difficult to assess because of a general lack of direct measurements with which to verify the model outputs. Nonetheless, it is generally accepted that these transport models only provide estimates of bedload transport (perhaps within 50 to 100% of the actual value). Gomez and Church (1989), for example, examined ten transport formulas and found that none of them were entirely adequate for predicting bedload transport in coarse-grained (gravel-bed) rivers.

5.2.2. Channelized Flow

The ability of a river to transport sediment or sediment-borne trace metals is governed by the balance between driving and resisting forces. The principal driving force is gravity (g), which is oriented vertically to the Earth's surface and equal to 9.81 m/s^2 . The component of this gravitational force that acts to move the available mass of water in the channel downslope is equal to $g \sin \beta$, where β is the average slope of the channel (Leopold et al. 1964). The resisting forces are more difficult to quantify and originate from a variety of sources. Attempts in recent years to identify these sources, and their relative importance, has met with only limited success. Nonetheless, the total resistance to flow can be broadly categorized into three principal components referred to as free surface resistance, channel resistance,

and boundary resistance (Bathurst 1993). *Free surface resistance* is caused by disruptions in the water surface by waves or abrupt changes in gradient. *Channel resistance* represents the loss of energy as water moves over and around undulations in the channel bed and banks, or is distorted by changes in the channel’s cross-sectional or planimetric configuration. The final component, *boundary resistance*, is produced by the movement of water over individual clasts or microtopographic features (e.g., dunes and ripples). Resistance produced by dunes and ripples is referred to as *form drag*, whereas resistance due to individual particles is called *grain roughness*. Other factors such as vegetation, suspended sediments, and internal viscous forces within the water also affect the resistance to flow.

Hydraulic engineers have been concerned with quantifying the relationships between the driving and resisting forces inherent in channelized flow for centuries. The net result of these studies has been the formulation of a number of resistance equations presented in Table 5.1. The Manning equation is perhaps the most widely used of these in the U.S., although the Darcy-Weisbach equation is growing in popularity within the academic community. The Manning equation was developed in 1889 by Manning in an attempt to systematize the available data into a useful form, and is expressed as:

(1)
$$V = (R^{2/3} S^{1/2})/n$$

where V is velocity, R is the hydraulic radius (i.e., the cross sectional area of flow divided by its wetted perimeter), and S is slope. The total resistance in the channel is defined by n, called the Manning roughness coefficient.

It is important to recognize that Manning’s roughness coefficient, as well as the other resistance factors, cannot be directly determined. Rather, it must be indirectly defined by measuring hydraulic parameters such as flow velocity, depth, and slope. For example, values of n are commonly determined by rearranging Eq. 1 and solving for n at a site in which the velocity, slope and hydraulic radius have been measured.

A common problem encountered in hydrological investigations is the need to calculate flow velocities for a range of conditions where roughness values have

Table 5.1. Selected flow resistance equations

Name	Equation	Unit notes
Manning equation	$V = k(R^{2/3} S^{1/2})/n$	$k = 1$ (SI units) $k = 1.49$ (imperial units) $k = 4.54$ (cms)
Chezy equation	$V = C\sqrt{RS}$	
Darcy-Weisback equation	$ff = (8gRS)/V^2$	

From Knighton (1998)
Symbols: C, n, ff – resistance coefficients; V – mean velocity; R – hydraulic radius; S – slope; g – gravitational constant

not been determined. Roughness values must therefore be estimated. This task is frequently undertaken by visually comparing the river or river segment being studied to other rivers of known roughness using some form of visual guide (e.g., Barnes 1968). Difficulties in using these visual guides arise because channel roughness varies as a function of flow depth and depends on factors, such as suspended sediment concentration, that cannot be easily accounted for using visual techniques. Needless to say, the method leaves a lot to be desired and estimated n values can lead to significant errors in the generated data. In fact, roughness estimates are one of the most important sources of uncertainty associated with numerical attempts at predicting sediment and contaminant transport rates.

The resistance equations presented in Table 5.1 show that the total resistance to flow is closely related to flow velocity. The velocity with which the water is traveling can vary vertically, horizontally, and downstream within the channel as well as through time at any given point. The variations in velocity that occur have led to various classifications of flow in open channels. For example, flow is frequently classified according to whether it is uniform or steady. Under *uniform flow* conditions, velocity is constant with respect to position in the channel; *steady flow* refers to the situation in which velocity is constant through time at a single point. Even a casual inspection of moving water in a river will reveal that the flow is inevitably non-uniform and unsteady. Flow can also be classified according to whether it is laminar or turbulent. In laminar flow, layers of water slide smoothly past one another without disrupting the path of the adjacent layers. Under this type of flow regime, the magnitude of the resistance is primarily governed by the molecular viscosity of the fluid, where viscosity is dependent on such variables as temperature. In contrast, turbulent flow is characterized by packets of chaotically mixing waters that transmit shear stresses across layer boundaries in the form of eddies. Eddies, or more correctly, eddy viscosity, greatly increases the resistance to flow and the dissipation of energy.

The transition between laminar and turbulent flow depends on both the flow depth and its velocity. The conditions at which the transition occurs can be predicted by the Reynolds number (Re), in which:

$$(2) \quad Re = VR(\rho/\mu)$$

where V is the mean velocity, R is the hydraulic radius, ρ is the density, and μ is the molecular viscosity of the water. In wide, shallow channels, R can be closely approximated by mean depth and, thus, they are frequently interchanged for one another. When values of Re are less than 500 flow is laminar; in contrast, values greater than 2,000 are indicative of turbulent flow. Flows possessing Re numbers between 500 and 2,000, called transitional flow, can exhibit characteristics of both laminar and turbulent flow conditions.

Flow in natural channels is turbulent, with the possible exception of a very thin layer of quasi-laminar flow along the channel bed, referred to as the *laminar sublayer*. Most turbulence is generated along the channel perimeter, causing the resistance to flow to increase, and the velocity to decrease, as the water-sediment

interface is approached. The maximum flow velocity, then, is generally observed immediately below the water surface near the center of the channel. However, the precise position of the maximum flow rates will vary with cross sectional shape as well as the channel’s planimetric configuration (Fig. 5.1).

5.2.3. **Entrainment**

A significant problem associated with the prediction of sediment transport is determining how the properties of flow in open channels combine to produce particle entrainment. *Entrainment* refers to all of the processes involved in initiating motion of a particle from a state of rest. It is clearly dependent on the erosive power of the flow. However, describing the relationship between the largest particle that can be entrained and the flow’s erosive power has proven difficult, because: (1) individual particles within the channel bed are acted upon by multiple forces, each of which is best described by a different parameter of flow, (2) flow velocities, particularly during floods, are neither constant, nor easily measured, and (3) the size, shape, and packing arrangements of particles within natural channels are highly variable,

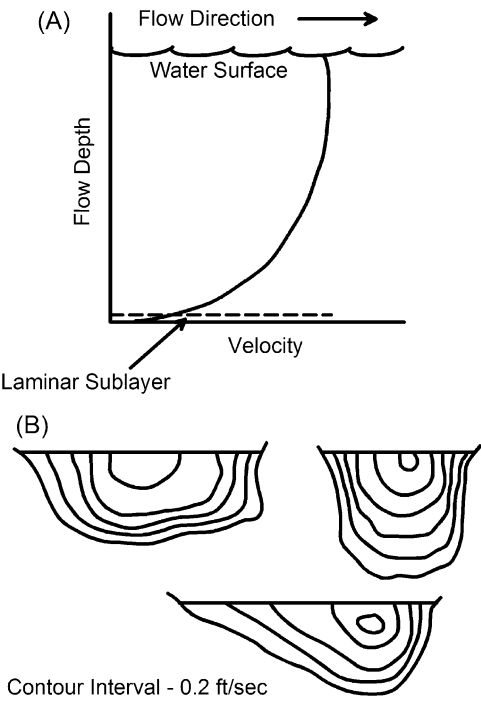


Figure 5.1. (A) Variations in flow velocity as a function of water depth. The laminar sublayer may be absent or discontinuous in coarse-grained rivers; (B) Typically observed variations in velocity through a cross section (Modified from Wolman 1955; Ritter et al. 2002)

and these parameters can cause divergent responses in particle motion to the same flow conditions (Ritter et al. 2002).

Historically, two methods have been utilized to predict a river's competence: critical bed velocity and critical shear stress. The adjective "critical" refers to the bed velocity or shear stress at the precise time at which the particle begins to move. Conceptually, the two approaches are quite different. Bed velocity refers to the impact of water on the exposed portions of the grain, and its strength is related to the momentum of the water (mass \times velocity). Tractive force or boundary shear stress (τ) is associated with the downslope component of the fluids weight exerted on a particle in the channel bed. In the form of an equation, it is expressed as:

$$(3) \quad \tau = \gamma RS$$

where γ is the specific weight of the fluid, R is hydraulic radius, and S is slope. In wide shallow channels such as those which typically transport gravel, water depth (D) is a close approximation of the hydraulic radius; thus, tractive force is proportional to the product of depth and slope.

Hjulström (1939) presented a set of curves that illustrates the relationships between velocity, particle size and process (Fig. 5.2). The dark, upper line on Fig. 5.2 represents the mean velocity above which a particle of a given size will be entrained. The lower dashed line indicates the approximate velocity at which the particle will be deposited. Hjulström's curves, then, illustrate that higher velocities

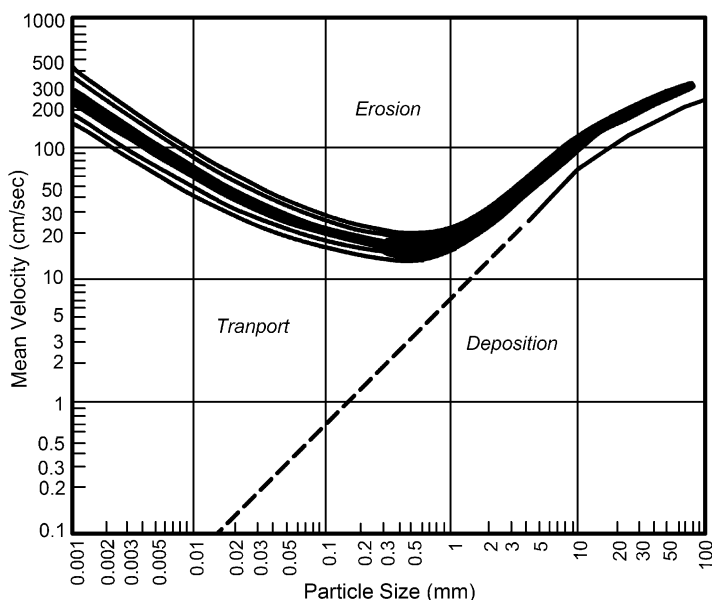


Figure 5.2. Mean velocity at which particles of various size are eroded, transported, and deposited (From Hjulström 1939, In Trask P (ed.) Recent Marine Sediments, American Association of Petroleum Geologists, used with permission)

are required to initiate the movement of a particle from a state of rest than to keep it in motion after it has been eroded. More importantly, the erosional threshold between grain size and velocity (upper set of lines) is not characterized by a linear trend in which continually increasing velocities are required to entrain larger and larger particles. Rather sand-sized particles between approximately 0.25 and 2 mm are the easiest to erode, and higher velocities are required to entrain particles that are both larger and finer than this size range. The higher velocities needed to entrain fine-grained (silt- and clay-size) particles can be attributed largely to the material's cohesive properties which bind the particles together. In fact, these fine-grained sediments tend to erode as aggregates rather than as individual particles (Knighton 1998), adding even more complexity to the problem of entrainment.

It is important to note that the Hjulström diagram is based on mean velocity and not critical bed velocities for the simple fact that the rate at which the water is moving immediately adjacent to the channel floor is extremely difficult to measure in high-energy rivers. In contrast, the depth and slope of the water in a channel are easily measured, making critical shear stress a more attractive parameter.

In the case of critical shear stress, the threshold for particle entrainment has been most notably illustrated by the Shields' diagram in which dimensionless critical shear stress (θ) is plotted against the grain Reynolds number. The grain Reynolds number is expressed as D_i/δ_o , where D_i is grain diameter and δ_o is the thickness of the laminar sublayer (Fig. 5.3). It essentially describes the extent to which an individual particle projects above the laminar sublayer into the zone of turbulent flow. Because D_i/δ_o is related to particle size, Fig. 5.3 is similar to the Hjulström diagram in that it depicts the force that is required to entrain a particle of a given size at the time of erosion. The Shields' diagram shows that the threshold of erosion reaches a minimum of approximately 0.03, which corresponds to particles in the size range of approximately 0.2 to 0.7 mm. The dimensionless critical shear stress required to entrain particles smaller than 0.2 mm increases because the particles reside entirely within the laminar sublayer where they are less likely to be affected

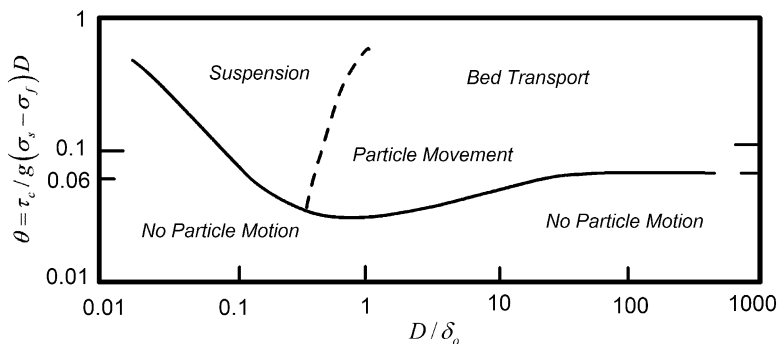


Figure 5.3. Shields' diagram for entrainment of bed particles. D is grain diameter, τ_c is critical shear stress, σ_s is sediment density, σ_f is fluid density, and δ_o is thickness of laminar sublayer (From Ritter et al 2002)

by turbulence. For particles larger than about 0.7 mm, which extend above the laminar sublayer, dimensionless critical shear stress (θ) increases before reaching a constant value, which has been shown to range from 0.03 to 0.06 (Komar 1989). If this constant value of dimensionless critical shear stress is substituted into its mathematical expression, and rearrange to solve for τ_c we see that:

$$(4) \quad \theta = \tau_c / [g(\rho_s - \rho_f)D_i]$$

$$(5) \quad \tau_c = \theta g(\rho_s - \rho_f)D_i$$

where ρ_s is the density of the particle, ρ_f is the density of the fluid, g is the gravitational constant, and θ is a constant of dimensionless shear stress (which falls within the range of 0.03–0.06).

Equation 5, and similarly developed mathematical expressions for critical shear stress, provide a powerful tool for assessing the flow conditions required for entrainment because the shear stress required to move a particle of a given size can be easily determined by assuming an appropriate value of θ .

A disadvantage of using either critical shear stress or critical bed velocity is that both approaches ignore the forces of lift on the particle. Lift is an upward directed force that can be created by the development of turbulent eddies on the downstream side of a particle, or by the establishment of a pressure gradient as a result of differences in flow velocities over and around a clast. In either case, lift has the potential to reduce the velocity or shear stress that is theoretically required to entrain a clast of a given size (Baker and Ritter 1975).

A slightly different approach that has been used to quantify sediment transport is to estimate a flow's stream power. Stream power is mathematically defined as:

$$(6) \quad \omega = \gamma QS$$

where ω is stream power, γ is the specific weight of the fluid, Q is discharge, and S is slope. If we consider the stream power that is available to transport sediment per unit width of the channel, then Eq. 6 becomes:

$$(7) \quad \omega/w = \gamma QS/w = \gamma(wdv)S/w = \gamma vdS = \tau v$$

where w , d , v represent the width, depth, and velocity of the flow, respectively. Equation 7 illustrates that stream power is a parameter that includes measures of both shear stress and velocity. As a result, it has been widely used to describe the sediment transport capacity of rivers.

Studies of particle entrainment in coarse-grained rivers have demonstrated that coarser clasts tend to be more mobile, and finer clasts less mobile, than would be predicted for sediment of uniform size, shape, and packing arrangements (Parker et al. 1982; Paola and Seal 1995). The importance of these characteristics on entrainment seems to rest with their influence on the exposure of individual particles to the overlying flow. For example, it is now clear that finer particles can be partially

hidden by larger clasts, thereby increasing the force required for entrainment of the smaller particles. Entrainment may also be complicated by the development of stratigraphy within the channel bed in which coarse-grained sediments overlie and bury finer grained sediment, a process that accentuates hiding effects (Paola and Seal 1995). These observations have led to a concept known as the *equal mobility hypothesis*. This hypothesis suggests that the effects of particle hiding and sediment layering are so significant that all of the clasts in the channel bed will begin to move at the same shear stress (Parker et al. 1982; Andrews 1983; Andrews and Erman 1986). In other words, equal mobility implies that the size of the particles that are in motion within a channel does not change with increasing discharge in gravel-bed rivers; rather all of the clasts will begin to move at the same time upon reaching some critical threshold of shear stress. Actually, equal mobility in the strictest sense of the concept is unlikely to occur in natural channels except, perhaps, under extreme flood conditions (Komar and Shih 1992). Nonetheless, there is little question that entrainment processes are more significantly influenced by the size, shape, and packing of the bed sediment than originally envisioned.

5.3. PROCESSES OF CONTAMINANT DISPERSAL

If it were somehow possible to completely understand the mechanisms controlling the dispersal of contaminated particles, we could precisely determine the geographical patterns of trace metal concentrations within the river and its associated landforms, making site assessments considerably easier and less costly. Unfortunately, the dispersal of sediment-borne trace metals is dictated by a host of parameters that interact in complex ways to produce what are often confusing spatial trends in trace metal concentrations. Five distinct, but interrelated, processes are commonly cited as the predominant controls on contaminant transport rates and dispersal patterns (Lewin and Macklin 1987; Macklin 1996). They include (1) hydraulic sorting, (2) sediment storage and exchange with the floodplain, (3) dilution associated with the mixing of contaminated and uncontaminated sediment, (4) biological uptake, and (5) geochemical remobilization or abstractions. As we will see, the importance of each process varies between rivers as well as between reaches of a given river, making the geochemical patterns in trace metal concentrations all the more difficult to decipher.

5.3.1. Hydraulic Sorting

Hydraulic sorting involves the partition of particles into discrete zones along the channel according to their size, density, and shape. The sorting process actually involves several different mechanisms, the importance of which may vary with both time and space. The sorting mechanisms include: (1) selective entrainment as discussed earlier, (2) differential transport, in which smaller particles move farther than larger clasts following entrainment (Fig. 5.4), and (3) selective deposition

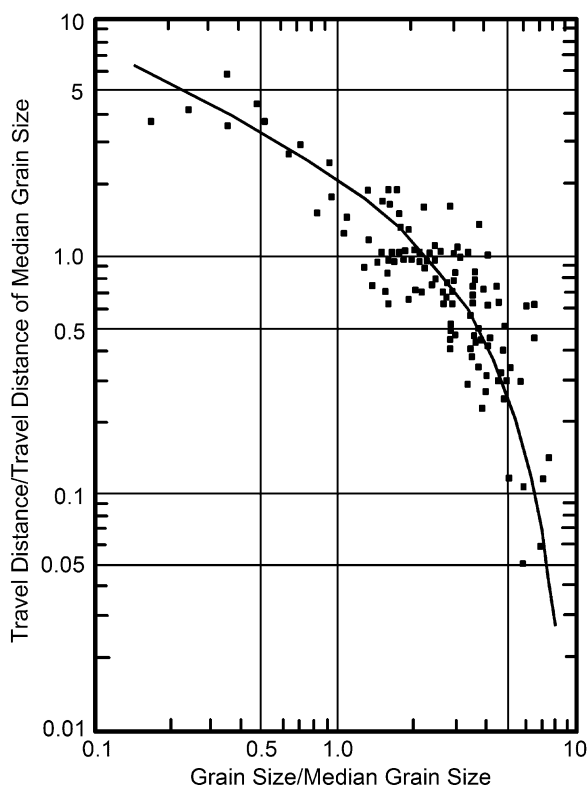


Figure 5.4. Change in travel distance as a function of grain size (After Hassan and Church 1992)

produced by differences in the settling velocity of the particles (Knighton 1998). The latter depends on the density and viscosity of water as well as the size, shape, and density of the sediment.

Later in the chapter we will see that hydraulic sorting can play a significant role in distributing trace metals within distinct morphological features in the channel, such as point bars and riffles. On a larger (basin) scale, it is primarily of significance where trace metals are associated with hydraulically heavier particles, such as metal enriched sulfides. In this case, the denser particles can be concentrated closer to the point source, and move more slowly downstream, than metals attached to hydraulically lighter particles which are transported entirely in suspension. For example, hydraulic sorting of sulfide minerals along the Rio Abaróá downstream of an impoundment containing mining and milling wastes is illustrated in Fig. 5.5. Here, hydraulic sorting produced a downstream decrease in both grain size of the sediment and the abundance of sulfide grains, thereby producing a minor downriver decline in metal concentrations.

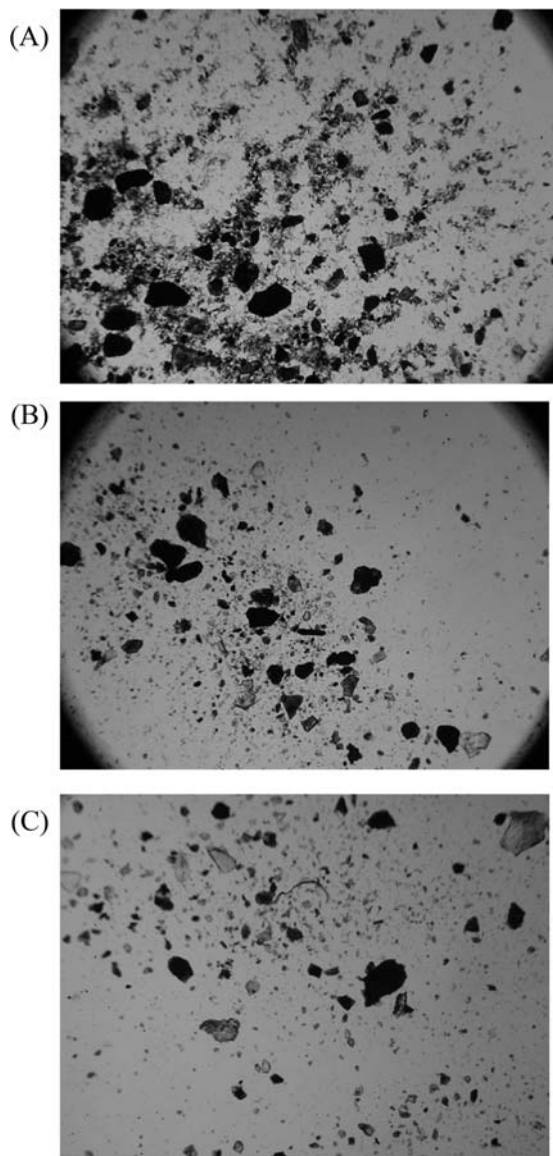


Figure 5.5. Photo micrographs of mill tailings at the Abaró Mine, southern Bolivia (A) and downstream channel bed sediments (B, C). Black grains are sulfide minerals (From Villarroel et al. 2006)

5.3.2. Dilution and Exchange with the Floodplain

A second important process influencing geographical patterns in trace metal concentration is the dilution of contaminated sediment by non-contaminated, or less

contaminated, sediment. The latter materials can be derived from tributaries or the erosion of the channel banks, floodplain, or some other geological feature. Dilution is one of the most commonly cited causes of decreasing downstream concentrations from a point source. It can occur gradually along the channel, or abruptly. Abrupt dilution is illustrated in Fig. 5.6 for the Rio Tupiza of southern Bolivia. Here Pb, Zn, and Sb concentrations decline upon mixing with the relatively clean sediments from the Rio San Juan del Oro. Dilution has less to do with the actual transport of contaminated particles than it does with altering the composition of the sediment load prior to its deposition.

Dilution can occur through three distinct processes. To illustrate their differences, let us assume that a sample of sediment from the channel bed of a river is contaminated with Pb and that all of the Pb is associated with silt- and clay-sized particles. One way in which this hypothetical sample can be diluted is by adding sediment of the same mineralogical composition and grain size, but which has a lower Pb concentration. In contrast, the material can also be diluted by adding sediment in which the concentration of Pb in the silt and clay is exactly the same as that in our contaminated sample, but the dilutant contains less fine-grained material (and, thus, possesses a lower bulk Pb concentration). Dilution, then, occurs as a result of a change in the grain size distribution of the original sediment. This type of dilution will be discussed later with regards to the use of grain size and compositional correction factors. While differences in the mechanisms of dilution are subtle, they can have significant effects on data interpretation. Take, for instance, the influx of sediment from a tributary in which the concentration of Pb in silt and clay sized particles is the same as that observed along the axial channel. Trace metal concentrations would appear unchanged downstream of the tributary if only the

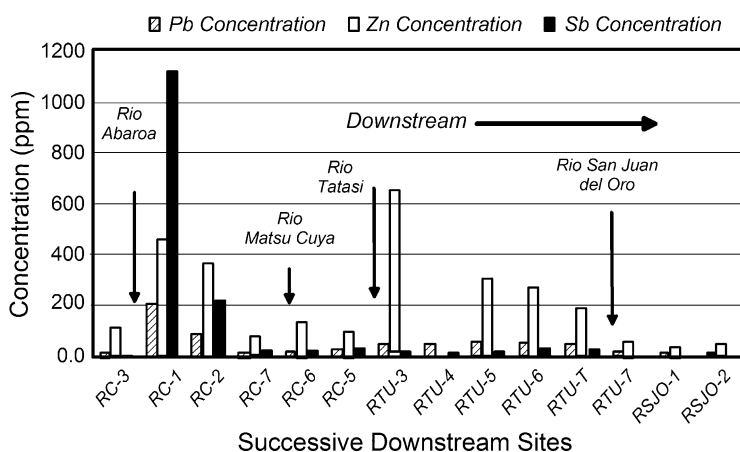


Figure 5.6. Concentration of Pb, Zn, and Sb in channel bed sediments of the Rio Chilco-Rio Tupiza drainage system, southern Bolivia. Antimony mines occur within the Rio Abaró and Rio Matsu Cuya basins; Polymetallic tin deposits are mined in the Rio Tatasi (From Villarreal et al. 2006)

< 63 μm sediment fraction was analyzed. However, concentrations downstream of the tributary could increase, decrease, or remain the same if we analyzed the bulk sample. Exactly which result will occur depends on the percentage of sand in the tributary and axial channel sediment, and the effects of dilution associated with their mixing.

In addition to the processes of dilution described above, dilution can also occur by dispersing a finite amount of contaminated sediment over a larger area. This form of dilution is particularly important following the removal of a contaminant source(s) by some form of remediation.

5.3.3. Sediment Storage and Exchange Mechanisms

Sediment storage and exchange with floodplain and channel bed deposits is often a significant control on contaminant transport rates and dispersal patterns. Storage can be thought of as the reverse of dilution. Rather than adding non-contaminated particles to the mixture, contaminated particles are removed from the load being transported by deposition upon the channel bed or floodplain, thereby decreasing their overall abundance.

The storage of sediment within the channel bed is closely linked to the reworking of bed sediments during flood events. Along many rivers the scour of the channel bed occurs during rising water levels as flow velocities and tractive force increase. The channel is subsequently filled during the waning stages of the flood (Fig. 5.7). During the scouring and filling episode contaminated particles are incorporated

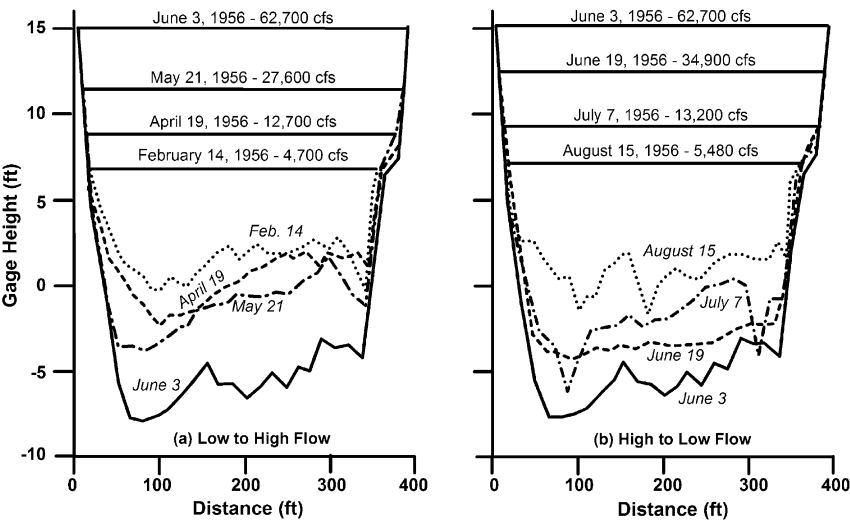


Figure 5.7. Scour and fill during the passage of a flood along the Colorado River at Lees Ferry, Arizona. (A) low to high flow; (B) high to low flow (From Leopold et al. 1964)

into the bed material where they are stored until they are remobilized by a subsequent high water event. It follows, then, that the amount of contaminants stored within the channel will depend, in part, on the degree to which the channel bed is eroded and refilled during floods, and the frequency of scouring and filling events. The significance of scour and fill tends to be particularly pronounced within ephemeral channels of semi-arid or arid environments. It is much more subdued within perennial rivers found in humid regions, and in coarse grained mountain streams which are armored by the winnowing of finer sediments during low flow (Leopold et al. 1964). Even where scour and fill is minimal, however, fine-grained, contaminated sediment can be deposited and stored between larger, relatively immobile clasts, or infiltrate into the channel bed materials.

Very few studies have attempted to quantify contaminant storage within the channel. The existing studies on the topic have shown that in-channel storage of both fine-grained sediment and trace metals is generally small. For example, in several studies conducted in the U.K., the storage of silt- and clay-sized particles to which trace metals are most likely attached was found to be less than about 10% of total suspended sediment load (Walling et al. 1998; Owens et al. 1999). Similarly, the storage of Pb and Zn in the Rivers Aire and Swale was less than about 3% of the total annual load (Walling et al. 2003). These data led Walling et al. (2003) to suggest that channel storage is of limited importance in regulating the downstream transport of contaminants through many river systems. It should be recognized, however, that the above data are for relatively stable rivers. In systems where aggradation has (or is) occurring, channel storage may be of more importance (Brookstrom et al. 2001, 2004; Villarroel et al. 2006). A case in point was provided by Brookstrom et al. (2001) who found that 51% of the Pb within the Coeur d'Alene River valley was associated with riverbed sediment. The main stem of the Coeur d'Alene was severely impacted by mine tailings and the storage of Pb within the channel was primarily driven by nearly 3 m of historic aggradation.

In most rivers, the amount of sediment and sediment-borne trace metals stored within the channel is significantly less than that found within the floodplain where more than 40–50% of the total annual load may be deposited (Marron 1992; Macklin 1996; Owens et al. 1999; Walling et al. 2003). The residence time of the contaminants also differs substantially between channel and floodplain deposits. Although poorly constrained, residence times for fine-grained channel sediments is generally less than 5 years (Walling et al. 2003); in contrast, contaminants may be stored for decades, centuries, or even millennia in floodplain environments (Macklin 1996; Miller 1997; Coulthard and Macklin 2003).

Contaminant storage downstream of a point source generally leads to downstream declines in metal concentrations. However, storage is a time dependent phenomenon and while it may initially remove contaminated particles from the river's transported load, its effects on contaminated particles may decrease or even become reversed in later years. This follows because sediment is not only deposited on floodplains, but eroded from them. Thus, once significant quantities of metal-enriched sediments have been deposited on (and stored within) a floodplain, the floodplain

itself can become a contaminant source to the river during bank erosion. When this occurs, longitudinal patterns in metal concentrations may more closely reflect the exchange of particles to and from the floodplain rather than their distribution from the original point sources. An example of the importance of storage processes on contaminant concentrations is illustrated by the Carson River system of west-central Nevada. During the late 1800s significant quantities of Hg-enriched sediments were deposited along the valley floor downstream of mill processing facilities. Although there is significant variation in the data, Hg concentrations within the valley fill generally decline downstream from the mills (Fig. 5.8a). However, Hg concentrations within the modern channel bed deposits do not decrease downstream from the mills, but quasi-systematically increase for approximately 25–30 km before subsequently declining (Fig. 5.8b). Miller et al. (1998) argued that the observed longitudinal trend resulted from the progressive erosion of dense Hg–Au and Hg–Ag amalgam grains stored within the valley fill, and their incorporation into the channel bed materials. Subsequent declines downstream were presumably related to dilution caused by the influx of clean sediment from a tributary, and the erosion of uncontaminated, pre-mining bank deposits which were more extensively exposed downstream.

5.3.4. Geochemical Processes and Biological Uptake

Contaminant dispersal patterns are not only controlled by physical processes, but also by biogeochemical processes including biological uptake. While the accumulation of trace metals by biota must theoretically influence sediment-borne trace metal concentrations, in most systems its effects are negligible. Other biogeochemical processes, however, can play a significant role. These trends are perhaps most evident in rivers affected by atypical Eh and pH conditions, such as systems impacted by acid mine drainage.

In rivers with typical Eh and pH conditions, sediment-borne trace metal concentrations can be significantly influenced by spatial variations in geochemical processes, even in rivers devoid of acid drainage. Hudson-Edwards et al. (1996), for example, found that in addition to hydrodynamic dispersal processes (i.e., dilution, hydraulic sorting, and storage), downstream declines in sediment-borne metal concentrations in the River Tyne were related to geochemical changes in sediment mineralogy which occurred during particulate transport. More specifically, thermodynamically unstable minerals, such as galena and sphalerite, oxidized to form secondary minerals (primarily Fe and Mn hydroxides) which contained proportionately less trace metals than the original mineral grains. Clearly, the chemical form (speciation) of the trace metals changed along the channel as well.

Evans and Davies (1994) also demonstrated that geochemical processes can affect the metal content of the sediment as well as their chemical form. In their study, Mn coatings within the River Ystwyth in mid-Wales adsorbed significant quantities of Pb within bedrock confined channel reaches. In contrast, most of the Pb along meandering reaches, bound by alluvial sediments, was associated with Fe-oxides

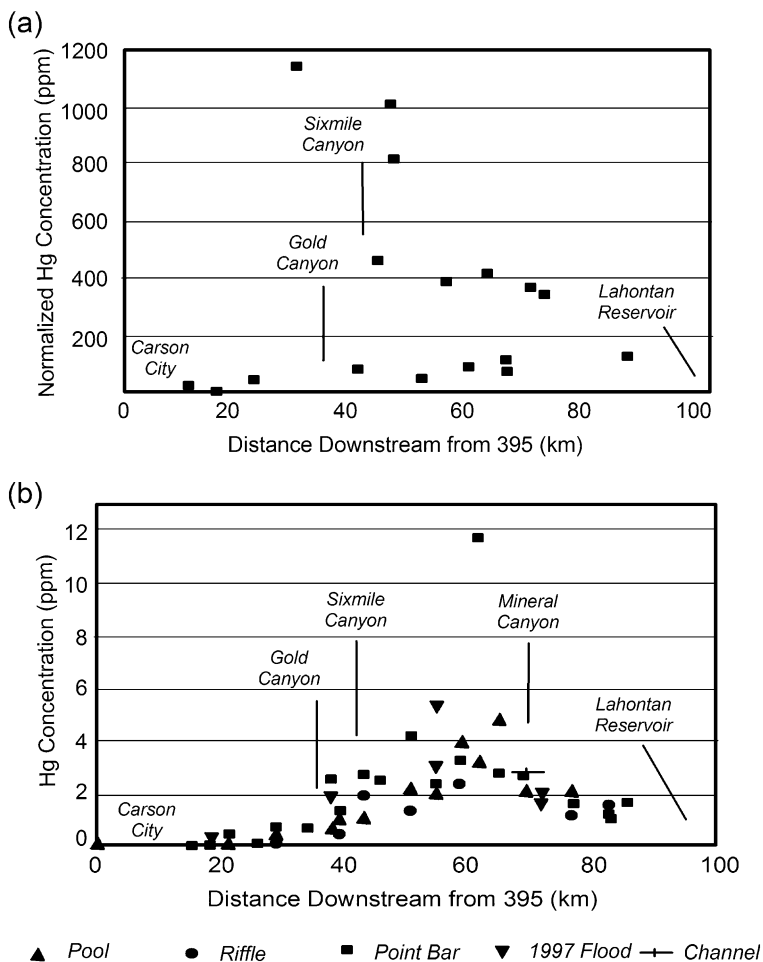


Figure 5.8. (A) Downstream changes in Hg concentration within valley fill located along the Carson River, Nevada. Elevated values are related to the release of Hg from ore processing mills which utilized Hg amalgamation methods; (B) downstream changes in Hg concentrations with the channel bed sediments. Note differences in concentration at a site between point bars, pools, and riffles, and the similarity in concentration between pre- and post-1997 flood deposits (Modified from Miller et al. 1998)

(and was therefore more tightly bound to the sediments). The observed differences in chemical speciation was attributed to the development of turbulent conditions within bedrock chutes that changed the oxygen status of the water enough to induce the precipitation of Mn(IV) onto the surface of sand sized particles. Thus, oxyhydroxide precipitation zones were created that were directly attributable to channel morphology and which impacted the chemical speciation of Pb along the river.

5.4. DOWNSTREAM PATTERNS

Every contaminated river exhibits its own unique downstream trend in elemental concentrations. Nonetheless, some recurring patterns are found to exist at the basin scale. In industrialized areas metal concentrations in rivers tend to increase semi-systematically downstream as they flow from relatively undeveloped headwater areas to more developed parts of the watershed, characterized by broad floodplains. Within the Aire-Calder River system in the U.K., for example, Pb, Cr, Cu, and other contaminants increase downstream as a result of urban and industrialized inputs to the mid- and lower reaches of the watershed (Owens et al. 2001; Walling et al. 2003) (Fig. 5.9).

Another frequently recurring pattern is associated with mining operations in mountainous terrains. In this situation, mining and milling debris, primarily from historic operations, is flushed into headwater tributaries and largely transported downstream until reaching the mountain front where the waste materials are deposited on agricultural lands. Metal concentrations tend to decrease downstream with distance from the mines, but deposition and storage of the contaminated debris may increase several fold upon exiting the mountain front.

Clearly, the above examples show that geographical patterns in concentration reflect points or zones of contaminant influx. Therefore, documenting the spatial patterns for a given metal allows us to identify potential sources of contamination even when multiple sources exist. The potential to identify contaminant

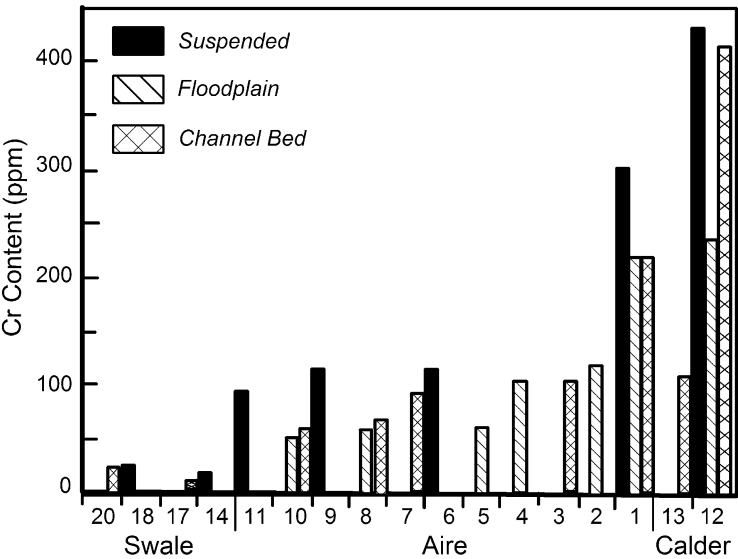


Figure 5.9. Downstream changes in Cr concentration in suspended, floodplain, and channel bed sediment of the Rivers Swale, Aire, and Calder. Numbers represent consecutive downstream sampling sites. Only the < 63 μ m fraction of floodplain and channel bed sediment was analyzed (From Owens et al. 2001)

sources is demonstrated in Fig. 5.6 for the Rio Chilco-Rio Tupiza drainage system of southern Bolivia. Increases in Pb, Zn, and Sb concentrations increase immediately downstream of tributaries affected by mining operations indicating that waste products from the mines are entering the axial drainage system. Moreover, the downstream patterns provide insights into the relative quantities of metal influx from each source, and differences in the composition of the mine wastes. The Rio Abaró, contaminated by an Sb mine, primarily provides Sb to the Rio Chilco. In contrast, the Rio Tatasi delivers large quantities of Pb and Zn to the axial drainage, and only limited quantities of Sb (Fig. 5.6).

Along most rivers, dispersal processes combine to produce a downstream decrease in metal concentrations from a point or zone of trace metal influx (Fig. 5.10). The downstream decay in metal concentrations can vary significantly from one point source to another, even along the same river. This too is well illustrated by the Rio Chilco-Rio Tupiza system. Between the Rio Abaró and the Rio Chilcobija concentrations rapidly decrease downstream, whereas trace metal content below the Rio Tatasi decays gradually until reaching the Rio San Juan del Oro where concentrations abruptly decline (Fig. 5.6). Moreover, the downstream declines in Sb differ from those of Pb indicating that the trends are metal specific.

Considerable attention has been given in recent years to documenting and modeling such trends in sediment-borne trace metal concentrations downstream of point sources. The first quantitative descriptions were provided by Wolfenden and Lewin (1977) using exponential decay and regression analyses. Since their seminal

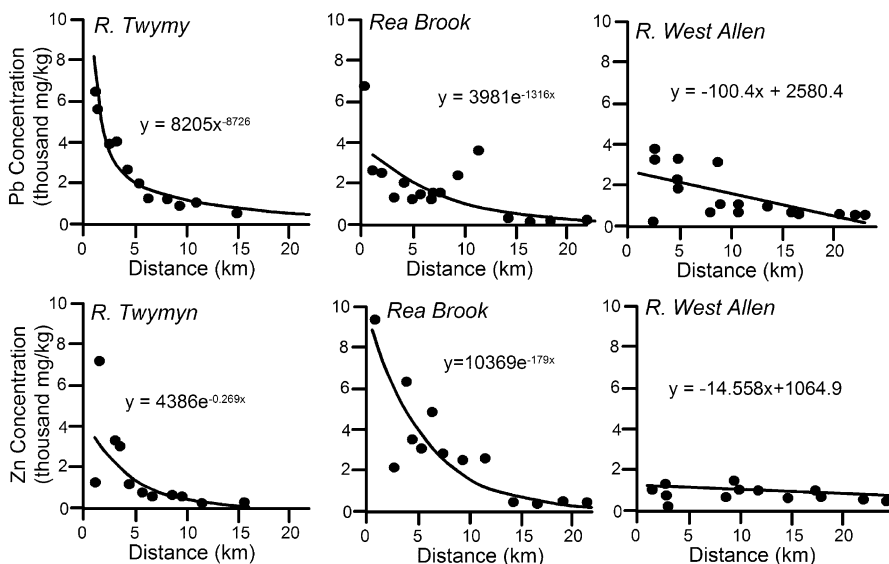


Figure 5.10. Downstream trends in Pb and Zn concentrations for three rivers in the U.K. Lines represent best-fit regression models (Modified from Lewin and Macklin 1987)

work was published, a number of other investigators have used the regression approach, applying linear, logarithmic, polynomial, power, and exponential decay models to a wide range of river systems (Lewin and Macklin 1987; Marcus 1987; Taylor and Kesterton 2002). Part of the interest in modeling these downstream trends has been the assumption that differences identified between specific trace metals for a given river provide clues regarding the transport rates and dispersal mechanisms. Comparison of the regression models, for example, has been used to identify downstream patterns in trace metal dilution, differential movement of metals by hydraulic sorting, and the relative importance of floodplain deposits as a source of sediment-borne contaminants (Macklin and Dowsett 1989; Macklin 1996). Care must be taken, however, in using regression equations to determine physical dispersal mechanisms because as we saw earlier in our discussions, dispersal is not always dominated by the physical transport and deposition of particles, but by geochemical processes (Macklin and Dowsett 1989; Taylor and Kesterton 2002). Where geochemistry prevails, the regression models may provide a basic understanding of metal concentrations between sampling sites, but they could potentially lead to erroneous conclusions as to why those concentrations exist.

Another less frequently used approach is to model the downstream dispersion of sediment-borne trace metals by assuming that concentration at any given point along the channel is related solely to the mixing of contaminated and non-contaminated sediments. The concepts for this approach were derived by exploration geochemists who recognized that downstream changes in trace metal concentrations were primarily driven by mixing of sediment from a geochemical anomaly with un-enriched tributary sediment. In the case of exploration, the area and grade of the anomaly can be predicted using readily available data including elemental content within axial channel sediments, basin area, and the background concentration of the element measured in tributaries. The model, however, can be reversed to predict the concentration of an element within the axial channel, provided that data are available concerning the size and grade of the upstream anomaly or contaminant point source (Helgen and Moore 1996). The revised formula is expressed as:

$$(8) \quad [C_a]_{km} = C_b \frac{(A_t - A_o)}{A_t} + C_o \frac{A_o}{A_t}$$

where $[C_a]_{km}$ is the concentration of the metal in the axial channel at a given distance from the point source, C_b is the average background concentration in the basin, A_t is the total basin area upstream of the sampled site, A_o is the area of the anomaly or point source, and C_o is the concentration of the metal in the surface anomaly. As written the model is most easily applied to natural geochemical anomalies. However, it has been modified and used by Helgen and Moore (1996) to describe downstream trends in concentration where the additional influx of trace metals occurs as a result of mining activity. They also demonstrated that a comparison of metal dispersion prior to and following mining activity can be used to characterize the degree of contamination along the channel and to set remediation targets.

The mathematical formula used by Helgen and Moore (1996) does not allow for variable rates of erosion and sediment influx from tributary channels. Marcus (1987), however, utilized a different form of mixing model to account for varying rates of sediment influx to the axial channel from tributary sources. In this case, downstream trends in Cu concentration along Queens Creek, Arizona were predicted on the basis of:

$$(9) \quad C_r = \left| \frac{(C_m)(X_m)}{(X_m)(X_t)} + \frac{(C_t)(X_t)}{(X_m)(X_t)} \right|$$

where C_r is the concentration downstream of the confluence along the axial channel, C_m is the concentration of the metal upstream of the confluence, C_t is the concentration of the metal within the tributary, and X_m and X_t are the basin areas or basin sediment yields of the axial channel and tributary, respectively. For Queens Creek, the mixing model performed better when compared against actual data than did developed regression models. The primary advantage of this particular mixing model is that it more accurately simulates the effects of sediment input on trace metal concentrations from tributaries characterized by variable sediment yields and trace metal contents.

Mixing models in general, however, are based on several simplifying assumptions that can limit their application in some rivers. The assumptions are that: (1) the downstream trends in concentration are the sole product of dilution resulting from the steady-state mixing with tributary sediment, (2) the terrain is characterized by uniform rates of erosion, either throughout a tributary catchment or over the entire basin, (3) there is only one source of metal-enriched sediment within the basin, and (4) the trace metals behave conservatively within the channel (i.e., they are not affected by geochemical processes). Perhaps the most common violations of the assumptions for contaminated rivers involves the influx of contaminated sediment stored in floodplain deposits during bank erosion, and the effects of processes other than dilution on trace metal dispersal (e.g., associated with hydraulic sorting or downstream variations in sediment storage).

Regardless of whether regression and mixing models are applied to the river, their fit to the actual data is often poor. For example, Macklin and his colleagues have intensively studied the River Tyne in the U.K. They found that at the basin scale, downstream trends in concentration could be adequately described using simple regression models, but when viewed at a finer scale, systematic variations in the data appeared. These variations, in part, could be attributed to increases or decreases in concentration immediately downstream of tributaries. More frequently, however, spatial changes in concentration formed a quasi-systematic wave like pattern that could not be attributed to tributary junctions. Macklin and Lewin (1989) argue that the underlying control on the observed pattern was the organization of the valley floor into alternating "transport" and "sedimentation" zones. The transport zones tended to be steeper and narrower than average, allowing sediment entering the reach to move more quickly through it with only limited storage. In contrast, the sedimentation zones possessed wider valley floors and shallower gradients

promoting deposition. Thus, it was along these reaches that most of the Pb and Zn from historic mining operations were deposited and where metal concentrations were the highest (Macklin and Dowsett 1989; Macklin 1996). Sedimentation zones have been recognized along many other river systems as well, such as the Rio Pilcomayo where deposition is closely linked to valley width (Fig. 5.11).

Data from the River Tyne also suggest that sediment-borne trace metals may move through the system as a series of sediment pulses, or slugs, complicating the pattern predicted by standard regression or mixing models (Nicholas et al. 1995; Macklin 1996). Sediment slugs are injections of clastic sediment into the channel as a result of both natural and human activities that exceed the transport ability of the river (Hoey 1992; Bartley and Rutherford 2005). The injected sediments result in cycles of aggradation and degradation which propagate downstream. Slugs, then, are associated with periods of disequilibrium that range from minor disturbances to basin-scale perturbations in sediment supply that may lead to a total readjustment of the valley floor (Table 5.2) (Gilbert 1917; Nicholas et al. 1995). With regards to metal transport, slugs are important because aggradation commonly increases sediment and contaminant storage and reduces their rate of downstream transport, while the subsequent period of degradation is likely to remobilize sediment associated trace metals, thereby increasing downstream transport rates.

5.4.1. Impoundment Failures

The transport processes described thus far have been related to “normal” rainfall-runoff events. However, waste disposal impoundments (ponds) associated with mining and milling operations are located along a large number of rivers, particularly in mountainous terrains. While the nature of these facilities is highly variable, the potential exists for the confining earthen structures to fail, catastrophically releasing



Figure 5.11. Transport and sedimentation zones along the Rio Pilcomayo, Bolivia

Table 5.2. Classification of sediment slugs

Slug size	Dominant controls	Impact on fluvial system
Macroslug	Fluvial process-form interactions	Minor channel change
Megaslug	Local sediment supply & valley-floor configuration	Major channel change
Superslug	Basin-scale sediment supply	Major valley-floor adjustment

From Nicholas et al. (1995)

metal enriched sediment into an adjacent stream channel. In fact, during the past 40 years, more than 75 major failures have released trace metals and other contaminants into riverine environments (Table 5.3). This equates to an average of nearly two major tailings dam failures per year, not including those in secluded regions, which are seldom reported.

Few studies have documented the resulting downstream trends in metal concentrations immediately after a failure (and before reworking by subsequent flood events). Graf (1990), however, found that the flood wave resulting from the 1979 Church Rock uranium tailings spill did not produce a systematic downstream trend

Table 5.3. Selected tailings dam failures

Mine (location – year)	Volume of material released – m ³
Aznalcóllar (Spain - 1998)	1,300,000 ^a
Harmony, Merries (South Africa – 1994)	600,000 ^b
Buffalo Creek (USA-1972)	500,000 ^b
Sgurigrad (Bulgaria – 1996)	220,000 ^b
Aberfan (UK – 1966)	162,000 ^b
Mike Horse (USA – 1975)	150,000 ^b
Bilbao (Spain – 1969)	115,000 ^b
Baia Mare (Romania - 2000)	100,000 ^c
Ages (USA – 1981)	96,000 ^d
Huelva (Spain – 1998)	50,000 ^b
Stancil, Perryville (USA – 1989)	39,000 ^b
Dean Mica (USA – 1974)	38,000 ^b
Arcturus (Zimbabwe – 1978)	30,000 ^b
Maggie Pie (UK – 1970)	15,000 ^b
Borsa (Romania – 2000)	8,000 ^c

Modified from MRF (2002)

^a Metal enriched tailings

^b Waste

^c Effluent & tailings

^d Coal refuse

^e Cyanide contaminated effluent & tailings

in ^{230}Th concentrations within an entrenched arroyo system in New Mexico. Instead, concentrations were inversely correlated to unit stream power and the length of time that shear stress exceeded critical values during the passage of the flood wave (Fig. 5.12). In other words, sediment enriched in ^{230}Th was preferentially deposited within low energy environments. Along reaches where unit stream power was relatively high, most of the contaminated sediment was transported downstream. As was the case for the River Tyne described above, the concentration of sediment-borne contaminants was closely linked to the existing morphology of the channel and valley system.

In contrast to morphologically controlled cases, the 1998 Aznalcóllar Mine spill in Spain produced a highly sediment-laden flow ($\sim 660\text{ g/L}$ in solid weight) that resulted in a semi-systematic decrease in thickness of deposited tailings on the floodplain, despite the fact that significant differences in channel form and slope were present (Gallart et al. 1999) (Fig. 5.13). Crevasses splays (described in more detail in Chapter 6) were, however, preferentially formed downstream of narrow reaches and channel bends (Gallart et al. 1999). Splays and other overbank deposits produced during subsequent floods between January and May, 1999 also exhibited a general downstream decrease in metal concentrations (Hudson-Edwards et al. 2003). Differences in the downstream transport of sediment-borne trace metals during dam failures such as at Church Rock and Aznalcóllar are clearly an area requiring additional study. Nonetheless, differences are related to the regional physiography, the morphology of the channel and valley floor below the impoundment, the rheology of the flow, the timing of the failure relative to rainfall-runoff events, and the amount and size of the waste material that is released, among others factors.

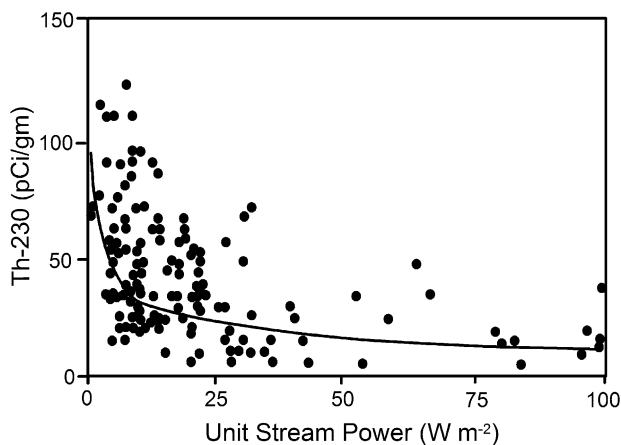


Figure 5.12. Change in ^{230}Th concentration with stream power, Puerco River, New Mexico (From Graf 1990)

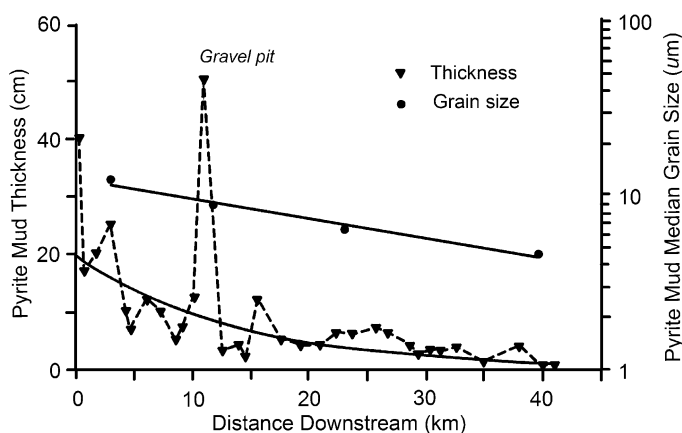


Figure 5.13. Change in the thickness and grain size of pyrite enriched mud deposits associated with the Aznalcóllar Mine tailings spill (From Gallart et al. 1999)

5.5. DEPOSITION AND STORAGE ALONG A REACH

Up to this point we have been concerned with the dispersal of sediment-borne trace metals over long reaches of the channel, often encompassing 10 to 100s of kilometers in length. However, basin scale trends in trace metal concentrations are the product of entrainment, transport, and deposition of particles at much smaller scales. The interaction of these geomorphic processes over the length of a reach (measured in 10's of meters) is of significant importance to site assessments because it can result in the concentration of trace metals within morphological features (e.g., bars, pools, and riffles) possessing distinctive sedimentological characteristics. In other words, the selective nature of specific particles sizes and densities by these morphological units is an extremely important, but often overlooked, consideration because it produces significant spatial variations in trace metal concentrations (Ladd et al. 1998; Taylor and Kesterton 2002). Failure to recognize these trends can lead to either over- or underestimation of the average (or maximum) concentrations that exist. In addition, sampling of channel bed sediments without regard to the depositional feature from which the materials were derived can confuse larger (basin) scale geographical patterns, thereby resulting in erroneous conclusions.

The morphological features found on the channel floor can be subdivided into small transitory deposits and larger (alluvial bar) deposits (Table 5.4). The transitory deposits are intimately tied to local flow conditions and fluctuations in those conditions through time. Larger, more persistent features, which represent our primary concern, are more closely related to discharge, sediment supply, and macroscale channel processes than local fluid hydraulics (Knighton 1998). These larger features are both created by and influence the mean flow configuration through the reach, and, therefore, have specific associations with a given channel pattern

Table 5.4. Characteristics of channel deposits

Scale	Characteristic
<i>Transitory deposits</i>	Bedload temporarily at rest
Micro-forms	Coherent structures such as ripples with λ ranging from 10^{-2} to 10^0 m
Meso-forms	Features with λ from 10^0 to 10^2 m; includes dunes, pebble clusters and transverse ribs
<i>Alluvial bars</i>	Formed by lag deposition of coarse-grained sediment
Macro-forms	Structures with λ from 10^1 to 10^3 m such as riffles, point bars, alternate bars, and mid-channel bars
Mega-forms	Structures with $\lambda > 10^3$ m such as sedimentation zones

Adapted from Knighton (1998), Church and Jones (1982) and Hoey (1992)

(e.g., point bars, pools, and riffles with meandering streams, alternate bars within straight channels, etc.) (Knighton 1998).

Historically, channel patterns were classified as straight, meandering and braided (Leopold and Wolman 1957). Most geomorphologists now recognize that this classification scheme is insufficient to describe the wide range of channel patterns that exist in nature. Schumm (1981), for example, combined the traditional scheme with the predominant type of load transported by the river to develop a classification system with 13 different patterns (Fig. 5.14). Others have subdivided rivers into those with single channels (straight and meandering) and those with multiple channels (e.g., braided and anabranching) (Knighton 1998). No matter which classification system is used, it is clear that the boundaries are indistinct. In fact, many rivers exhibit a single channel pattern during low flow conditions, but during floods possess a distinctly braided configuration. Thus, a river may acquire morphological features characteristic of both meandering and braided systems. Nonetheless, river classification is a useful process in that it aids in the description of the formative processes associated with the various planimetric configurations, and illustrates the general trends in channel pattern with changes in hydrologic and sedimentologic regime (Fig. 5.14).

The literature on channel patterns and their deposits is voluminous and complicated by an inconsistent use of terminology (see Church and Jones 1982, for a discussion). As such, an in-depth discussion is beyond our purposes here. Instead, we will focus on developing an understanding of the most important formative processes, and the basic characteristics of the resulting fluvial deposits.

5.5.1. Channel Patterns

5.5.1.1. Straight and meandering channels

In addition to possessing relatively linear banks, most straight channels exhibit accumulations of sediment, called *alternate bars* that are positioned successively downriver on opposing sides of the channel (Fig. 5.15a). Opposite the alternate bars are relatively deep areas called *pools* which are separated by shallower and

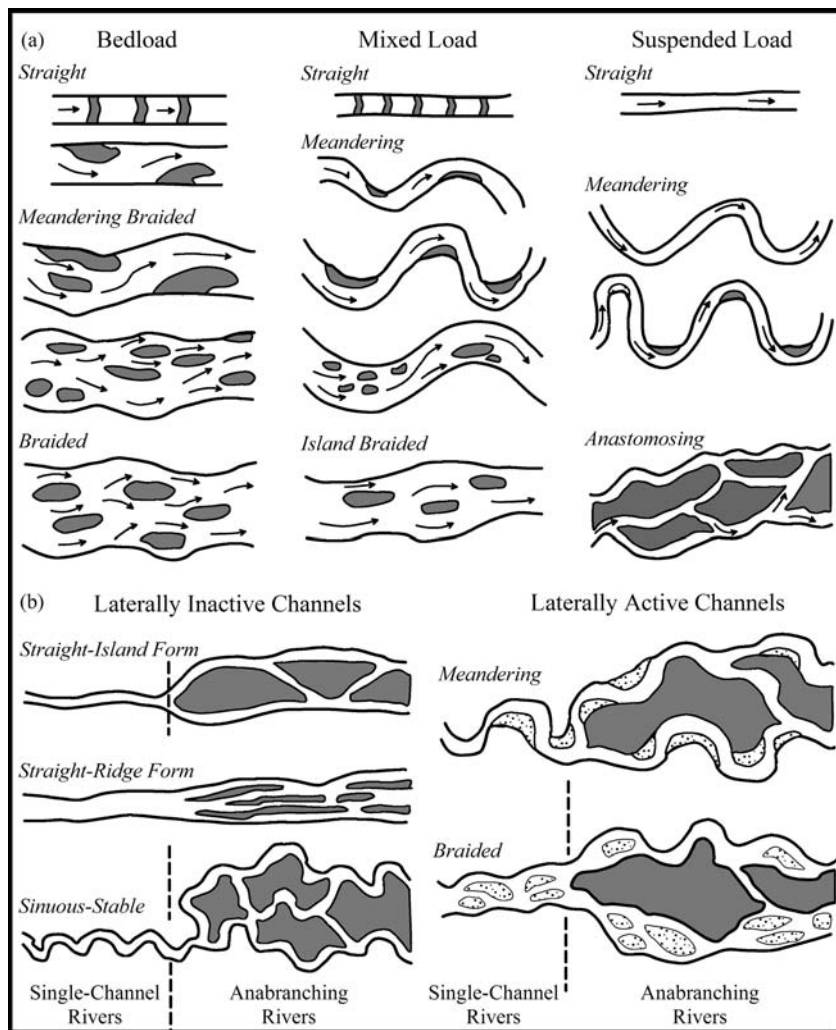


Figure 5.14. (A) Classification of channel pattern by Schumm (1981, 1985); (B) classification of anabranching channels adapted from Nanson and Knighton (1996) (From Huggett 2003)

wider reaches called *riffles* (Fig. 5.15c). The sequence of alternate bars, pools, and riffles is organized in such a way that the deepest part of the channel, referred to as the *thalweg*, migrates back and forth across the channel floor. A straight channel, then, possesses neither a uniform streambed nor a straight thalweg. Rather, the configuration mimics the sequence of features found in meandering rivers (Fig. 5.15b). The distinction between the two patterns (meandering and straight) is normally defined by a sinuosity of 1.5, where sinuosity is the ratio of stream length to valley length. The designated value of 1.5 is arbitrary and has no specific

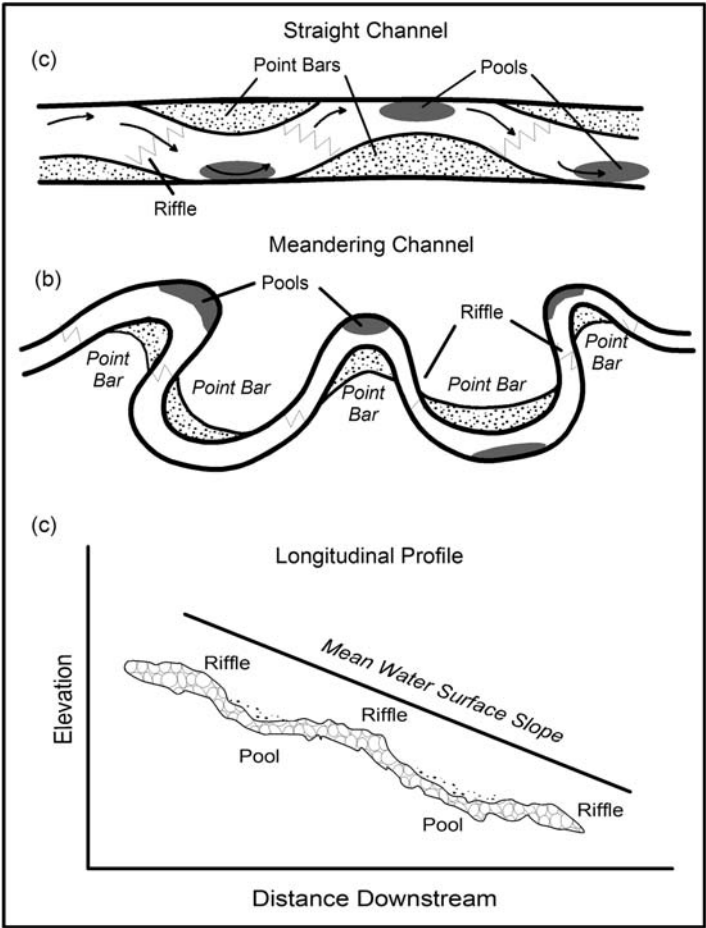


Figure 5.15. Planform morphology of straight (A) and meandering (B) channels. Longitudinal profile along thalweg (C)

mechanical significance. Channels with a sinuosity of 1.4 will likely exhibit similar flow patterns and sediment transport mechanics as channels possessing a sinuosity of 1.6 provided that all the other factors controlling channel form are the same.

Most rivers do not possess straight channels for long distances. In contrast, the meandering pattern is the most commonly observed planform. An important component of flow in meandering rivers is a secondary pattern of circulation that is oriented oblique to the downstream flow direction. Presumably, as water flows around a meander, centrifugal force leads to a slight elevation in the water surface on the outside of the bend (Fig. 5.16). The increase in elevation produces a pressure gradient that gives the flow a circular motion. This corkscrew type motion, referred to as *helical flow*, has traditionally been thought of as a single rotating cell that

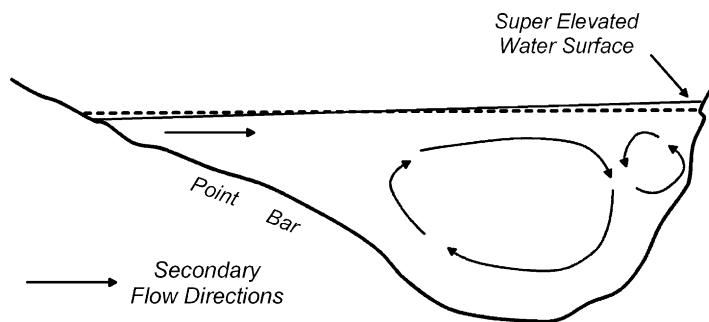


Figure 5.16. Secondary flow directions at a meander bend (From Markham and Thorne 1992)

reaches its maximum velocity within pools, immediately downstream of the axis of the meander bend. More recent studies have shown, however, that secondary circulation at meanders is more complex and may consist of several interacting, cells (Markham and Thorne 1992). Near the outer bank, the water interacts with bank materials causing flow to move upward along the channel margins before moving toward the center of the channel at the water surface (Fig. 5.16). The mid-section of the channel is characterized by the classical form of helical flow and may contain as much as 90% of the discharge (Markham and Thorne 1992). The flow direction of this cell is opposite to that occurring along the outer margins. As a result, the two cells meet to form a zone of flow convergence that reaches its maximum velocity along the base of the banks and at the channel bed, thereby promoting scour in these areas. Along the inside of the meander bend, there is commonly a component of flow that moves over the top of the point bar toward the opposing bank (Dietrich and Smith 1983; Dietrich 1987).

The direction of rotation of the secondary flow cells is reversed between successive meanders because the features within the channel are opposite to one another (Fig. 5.17). Originally, it was proposed that the reversal occurred between a pool and the next downstream riffle, where flow conditions promote channel bed deposition. However, the nature of the secondary currents existing between the pools appears to be more complex than originally thought. In fact, different flow patterns have been observed in different rivers for reaches located between two successive pools (Hey and Thorne 1975; Thompson 1986; Dietrich 1987; Markham and Thorne 1992). The observed differences in secondary circulation between rivers are apparently related to variations in channel planform, width/depth ratio, and flow stage (Dietrich 1987).

The pattern of pools, riffles and bars in meandering rivers is a manifestation of how flow, sediment transport and bedforms are interrelated. Successive riffles (or pools), for example, are usually spaced at distances of five to seven times the channel width. The spacing of the pool and riffle sequence is largely independent of the material forming the channel perimeter and, therefore, is thought to be predominantly related to larger scale flow patterns in meandering rivers. The sedimentology

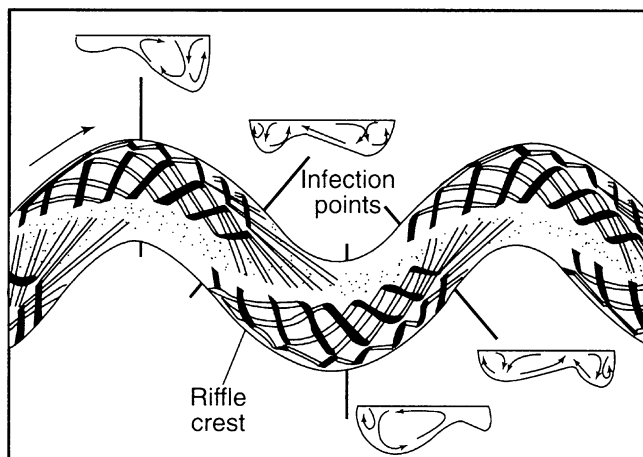


Figure 5.17. Model of secondary flow cells and flow lines in a meandering river (From Thompson 1986)

of the features also depends on the hydrologic conditions within the channel. At low flow, riffles tend to possess coarse grained sediments and are characterized by rapid, shallow flows and steep water surfaces; in marked contrast, the surface materials in pools are relatively fine grained, a trait consistent with the deeper, gentler flow conditions.

A significant question that has received considerable attention is how the pool and riffle sequence is maintained. Based on the flow patterns observed between periods of flood, one would expect the entrainment of the sediment at the coarse grained riffles to be transported downstream to the pools until the pools are destroyed by aggradation. Field measurements have shown, however, that the flow velocities during a runoff event increase at different rates between the pools and riffles (Fig. 5.18). These differences lead to a reversal in the flow rates so that the velocity within the pool is greater than that over the riffle during periods of high water. As a result, large particles eroded from the riffles can be transported through the pools during floods. During the recessional phase, the larger particles are deposited first on the riffle and the finer particles, eroded from the riffles, are redeposited in the pools (Fig. 5.18).

More recent studies have shown that the changes in velocity over pools and riffles with rising discharge may not be universally applicable to all rivers and that other mechanisms may account for the observed sedimentological and morphological patterns (Carling 1991; Carling and Wood 1994; Thompson et al. 1996, 1998). Nonetheless, there is little question that the competence of the flows within the pools must exceed that of the riffles during floods in order to maintain the pool and riffle sequence. In addition, it appears certain that the pool and riffle sequence is maintained by relatively large discharge events.

The scour and fill processes associated with the formation of pools and riffles can lead to the accumulation of channel lag deposits. These deposits are composed

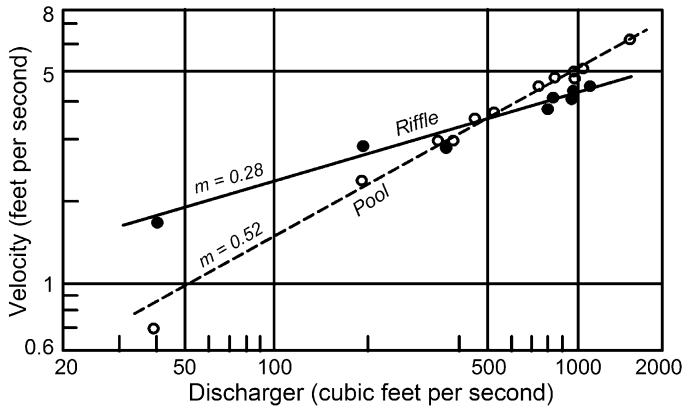


Figure 5.18. Change in velocity over a pool and riffle with increasing discharge, East Fork River, Wyoming (From Andrews 1979)

of coarse sediments that can only be moved during periods of flood. Typically the materials consist of coarse gravels, waterlogged plant debris, and consolidated aggregates and clasts of mud and clay from the eroded cut bank (Boggs 2001). Lag deposits are generally quite thin and laterally discontinuous. In fact, in rivers devoid of significant coarse sediment, they may be completely lacking. As we will see later in our discussion, channel lag deposits can contain accumulations of heavy minerals enriched in trace metals, particularly within rivers contaminated by mining debris.

During periods of low flow, channel lag gravels are likely to be buried by finer sediments which merge into the base of a downstream point bar. Point bars are the most significant features in meandering rivers. In cross section, they are characterized by a nearly horizontal surface on the inside of the meander bend that is positioned at the elevation of the adjacent floodplain. It then slopes gradually towards the thalweg until merging with the channel bed (Fig. 5.19). The point bar surface is the site of sediment deposition which occurs during the migration of the point bar across and downvalley (Fig. 5.20a). In the ideal case, decreasing velocities and flow depths over the point bar produce an upslope decrease in grain size that is accompanied by changes in bedforms (see Chapter 6). The actual character of the point bar, however, will depend on the size and composition of the sediment in transport. In rivers carrying a mixture of coarse and fine material, the upward progression is from basal gravels to sand to silt. The fine-grained silt cap at the top of the point bar is produced during high flow events by overbank deposition (Reading 1978). In rivers carrying fine-grained sediments, the upward transition often ranges from a fine sand layer near the base of the point bar to silty or clayey sediments near the top (Reineck and Singh 1980).

The ideal point bar rarely occurs in nature except perhaps along small meandering streams. A common complication to the ideal sequence is the flow of water across the point bar during floods which produces a pronounced channel called a *chute*. Chute channels tend to be accentuated on the upstream end of the point bar and

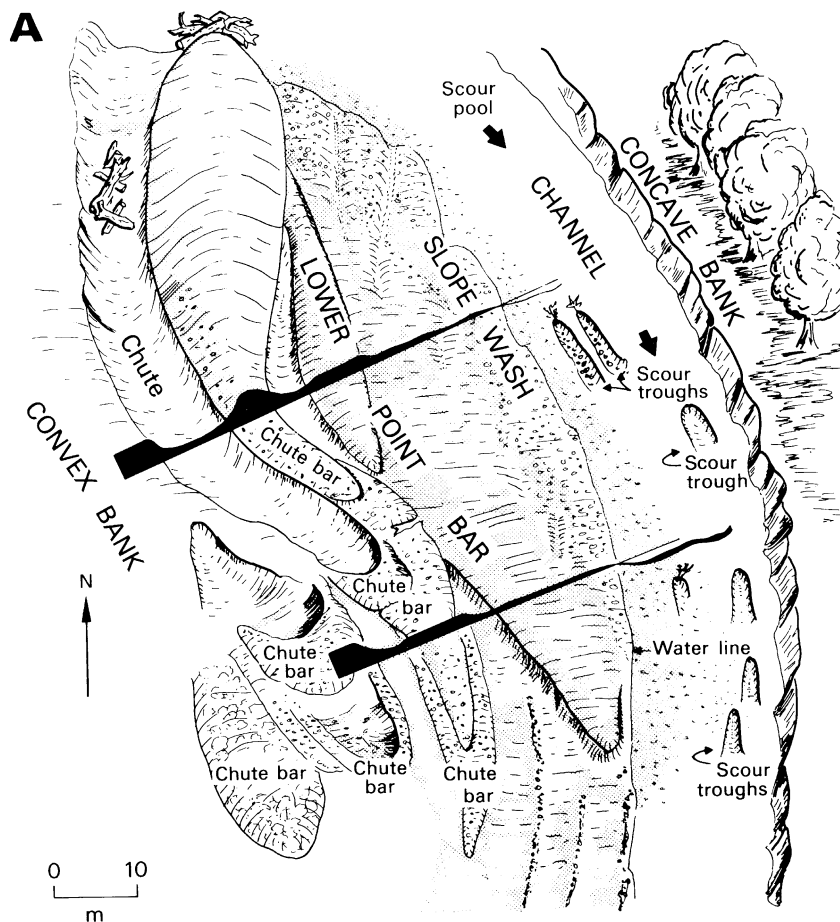


Figure 5.19. Topographic and sedimentology features of a coarse-grained point-bar (Original figure after McGowan and Garner 1970; reproduced from Reading 1978)

decrease in depth downstream, ultimately terminating in a bar composed of fine-grained sediment at the end of the chute (Fig. 5.19). In other cases, the chute may continue to erode, progressively capturing a larger portion of the flow. Eventually, flow diverted into the chute may completely dissect the point bar, resulting in the formation of a bar that subdivides the reach into two channels.

5.5.1.2. Braided rivers

The braided pattern is characterized by the subdivision of a single channel into a network of branches by islands of sediment (Fig. 5.20b) generically referred to as *braid* or *medial bars* (Fig. 5.21). In comparison to meandering rivers, braided channels tend to be highly dynamic; both the position and total number of channels

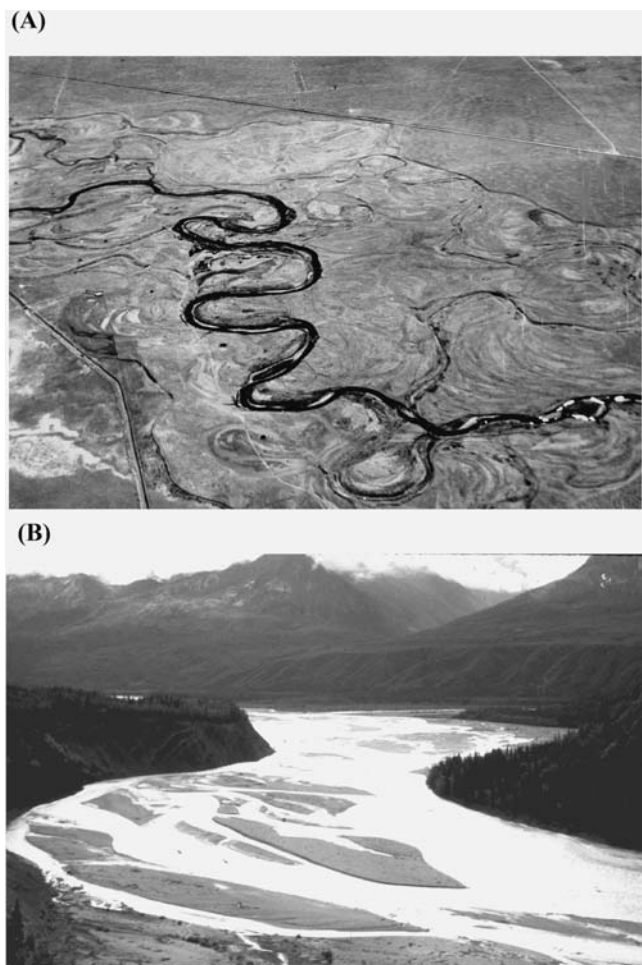


Figure 5.20. (A) Laramie River, Wyoming (photo by J.R. Balsley); (B) Matanuska River, Alaska (Photo by D. Germanoski)

within a braided reach can change significantly over a period of days. In addition, bars tend to be migratory and transient, inhibiting the development of a stable pool and riffle sequence (Church and Jones 1982). The dynamic nature of the braided system suggests that the residence time of contaminants stored within the channel bed will likely be shorter than within either point or alternate bars unless the river is aggrading (although data adequately supporting this contention is lacking).

Bar morphology and sedimentology in braided rivers is highly variable and depends in large part on the size and size distribution of sediment within the channel. Thus, braided rivers are commonly classified as either being coarse or fine-grained systems, while braid bars can be grouped into three basic types: longitudinal

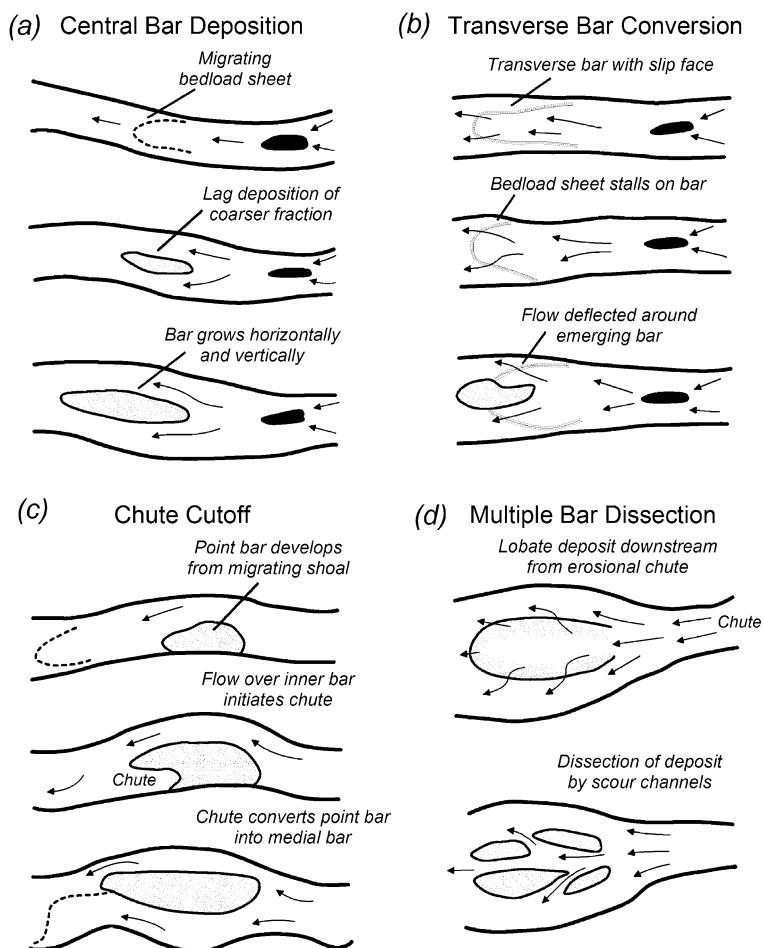


Figure 5.21. Mechanisms of braid bar development (From Knighton 1998)

bars, transverse bars, and lateral bars (Miall 1996; Boggs 2001). As the name implies, longitudinal bars are oriented roughly parallel to the general flow direction. Transverse bars possess avalanche faces oriented at an angle to stream flow (Boggs 2001). They are similar to subaqueous linguoid bars which exhibit a lobate or rhombic configuration and which resemble large dunes that migrate along the channel. Both linguoid and transverse bars are predominantly found in sandy rivers (although linguoid bars are perhaps more common in rivers with high bedload transport rates), and they are often associated with the capture of other migrating bedforms (Fig. 5.21) (Germanoski 2000; Boggs 2001). Lateral bars are attached to the margins of the channel and can be much larger than the other bar forms. Their formation is generally associated with the inability of the stream flow to transport

sediment through low energy zones located along the channel banks. Growth of lateral bars primarily occurs by lateral accretion similar to that associated with point bars (Miall 1992). Thus, they can exhibit a fining upward sequence, particularly within sandy rivers.

Detailed studies have shown that where braided and meandering reaches occur along the same river, the divided segments typically possess steeper gradients, shallower flows, and greater total channel width, although the width of the individual channels may be less than that of the undivided reach (Fahnestock 1963; Smith 1970, 1974). The classical model of braid bar formation, presented by Leopold and Wolman (1957), is referred to as the Central Bar Theory because the process begins with the accumulation of sediment near the center of the channel during high flow conditions (Fig. 5.21). This local accumulation of sediment is likely to become an incipient bar since the re-entrainment of the particles requires a higher velocity than does either their transport or deposition (Fig. 5.2). As sediment moves through the reach, some particles will be deposited at the downstream end of the incipient bar where water depth dramatically increases and flow velocities decline. Continued bar growth decreases the total cross sectional area of the channel until it is no longer capable of containing the entire discharge. Flow is then diverted around the bars causing erosion of the bank materials. The channel bed may also be deepened by scour. The combined effect of bed and bank erosion is the lowering of water level, a process that allows the bar to emerge as an island separating two channel branches.

Braid bars are also formed in many instances by the formation and subsequent dissection of bedforms (Fig. 5.21). Examples of these formative mechanisms can be found in Ashmore (1991), Ferguson (1993), and Germanoski (1990, 2000). Other mechanisms of bar formation have little to do with bedforms or in-channel deposition of sediment, but involve the cutoff of point and alternate bars (Fig. 5.21c; Ashmore 1991), and the reoccupation of abandoned channels (Eynon and Walker 1974; Germanoski 2000). Thus, the braided configuration can be formed by multiple processes which may function simultaneously along any given braided reach, although one specific mechanism is likely to be dominant. In all instances, the formation of a braided river is promoted by erodible banks, abundant bed load which can be temporally stored in bars, and rapidly fluctuating discharge (Fahnestock 1963). Perhaps the most important of these is the occurrence of erodible banks; if erosion is prohibited by cohesive bank materials or vegetation it is unlikely that braid bars will form. Fluctuating discharge, on the other hand, is not essential as laboratory studies have produced the braided pattern under constant discharge conditions. Nonetheless, there is no question that variations in discharge enhance cycles of erosion and deposition of sediment which is an integral part of the braiding process.

5.5.1.3. *Anabranching channels*

Nanson and Knighton (1996) defined a anabranching river as an interconnected network of channels separated by relatively stable alluvial islands that subdivide

the flow at discharges up to bankfull (Fig. 5.14). Perhaps the most common form is the anastomosing channel which is characterized by low gradients, low width/depth ratios, and fine-grained bed and bank sediment (Smith and Smith 1976; Smith 1986). Not all anabranching rivers, however, possess low-gradients or fine-grained particles (Miller 1991; Knighton and Nanson 1993). Nanson and Knighton (1996) defined six different types of anabranching channels ranging from low-gradient, fine-grained systems to steep, gravel bed rivers.

Anabranching rivers are rather uncommon and, until recently, the processes involved in their formation have received little attention. In fact, the mechanics of how the channel network is developed is still not fully understood, although it is now known that the network can be created by two fundamentally different processes. In one instance, channels are formed by avulsion in which channel bed aggradation leads to overbank flooding. The overbank flows cut a new channel into the existing floodplain deposits or scour out and reoccupy an abandoned channel. The other process involves the deposition of a ridge of sediment within the channel which subsequently becomes stabilized by vegetation and diverts the flow into two directions. Both of these formative mechanisms are promoted by the occurrence of: (1) stable, cohesive bank sediments that limit channel widening, (2) a hydrologic regime characterized by frequent, large floods, and (3) one or more mechanisms (e.g., ice or log jams, or channel sedimentation) which produces localized flooding (Nanson and Knighton 1996).

Although anabranching channels were once considered to be a type of braided river, it is important to recognize that the individual channels may acquire a straight, meandering, or braided configuration (Fig. 5.14). Thus, the sedimentology of anabranching channel deposits can broadly be described by the depositional models put forth for the other channel patterns. Deposits preserved within the floodplain sequence can, however, be distinctly different. We will examine these differences in the following chapter.

5.5.2. Trace Metal Partitioning Mechanisms

5.5.2.1. Grain size and compositionally dependent variations

Earlier it was argued that distinct morphological units within a river can possess significant differences in geochemistry. These differences are probably controlled by the types of features present and their sedimentological characteristics. For our purposes, it is important to understand what factors actually produce the observed variations in concentration. Perhaps one of the most important factors is the grain size distribution of sediment that comprises the morphological units of any given channel. The chemically reactive nature of silt- and clay-sized particles suggests that those containing an abundance of smaller particles should exhibit higher trace metal concentrations. In fact, grain size dependent variations in trace metal concentrations between morphological units have been recognized for a number of rivers. Ladd et al. (1998), for example, examined the variability in metal concentrations within and between seven different morphological units within a 500 m reach of Soda Butte

Creek, Montana. Five units were defined according to the nomenclature put forth by Bisson et al. (1982) for gravel-bed streams. They include lateral scour pools, eddy drop zones, glides, low gradient riffles ($< 1\%$ slope), and high gradient riffles ($1\text{--}4\%$ slope). Two types of bars were also defined according to Church and Jones (1982); they were attached bars and detached island bars. For all 12 of the metals examined, significantly different metal concentrations in the < 2 mm sediments were observed between the units. Eddy drop zones and attached (lateral) bars, which possessed the largest percentages of fine sediment, exhibited the highest concentrations of metals on average. In contrast, units with the largest quantity of coarse grained sediment, including glides, low gradient riffles, and high gradient riffles, tended to possess the lowest metal concentrations. Similar results have been derived for braided rivers, such as the Gruben located in the extreme arid environment of Namibia (Taylor and Kesterton 2002).

5.5.2.2. *Density dependent variations*

Not all of the differences in metal concentrations between the morphological units can be attributed to grain size, indicating that other factors must also be important. One of these factors is undoubtedly the association of metals with relatively dense particles which allows for their accumulation in high, rather than low energy environments. An example has been presented for the Carson River, Nevada by Miller and Lechler (1998). The Carson River was severely contaminated by historic mining operations which utilized Hg to extract Au and Ag from the Comstock Lode, one of the richest Au and Ag producing ore bodies in history. They found that point bar deposits exhibited higher Hg concentrations than the adjacent pools and riffles at any given site. The observed differences in concentration were attributed to the accumulation of sand-sized Hg–Au and Hg–Ag amalgam grains in point bars in the form of modern-day placer deposits.

Slingerland and Smith (1986) define a placer as “a deposit of residual or detrital mineral grains in which a valuable mineral has been concentrated by a mechanical agent,” in our case, running water within the river. The concentrated minerals generally possess a density much greater than quartz, such as gold, diamonds, cassiterite, or platinum group elements, and tend to be relatively durable so that they are not broken down during repeated long-term reworking by the river. While sulfide minerals are denser than quartz, they tend to breakup and decompose in oxygenated waters and, thus, rarely form placers in natural (unaltered) environments (Guilbert and Park 1986). However, in rivers contaminated by mining wastes, the sheer quantity of metal bearing sulfides released to a river may allow for their concentration in the form of a contaminant placer. A *contaminant placer* is defined here as a concentration of metal enriched particles by the hydraulic action of the river. Where they occur, trace metal concentrations will be locally elevated in comparison to other areas.

The concentration of heavy mineral grains from a heterogeneous mixture of sediment was originally thought to be created primarily by the more rapid settling of hydraulically heavier particles from the fluid. However, in a review of the processes

involved in placer formation by Slingerland and Smith (1986), it becomes apparent that particle sorting through settling is probably not as important to placer formation as originally believed. Rather it involves three processes described earlier, including selective entrainment, differential transport, and selective deposition. Interactions between these sorting processes can form placers over a wide range of spatial scales ranging from zones of abrupt valley widening to point bars to migrating dune foresets (Table 5.5) (Slingerland and Smith 1986). We are primarily concerned here with placers concentrated at the immediate scale that are associated with distinct morphological features of the channel. At this scale, placers can be categorized as those associated with bedrock channels and those formed in alluvium. Bedrock placers tend to develop within crevasses that can trap denser particles (e.g., joints, fractures, and faults), or in pools behind obstructions to flow (e.g., outcrops of resistant strata or dikes) (Fig. 5.22a). Common sites of placer formation in alluvial channels, at the intermediate scale, include the head of mid-channel bars or islands, the basal zones of point bars, and tributary junctions (Fig. 5.22b). Depending on the environment, point bars can be particularly important because the concentration of heavy minerals during channel migration can lead to the development of pay streaks, or in our case hotspots, within the associated floodplain (Fig. 5.23).

Smith and Beukes (1983) argued that placer formation associated with bars and tributary junctions can be attributed to stable convergent flow patterns that

Table 5.5. Observed sites of water-laid placers organized by spatial scale

Site/Scale	Site/Scale
Large Scale (10^4 m)	Intermediate scale (10^2 m)
Bands parallel to depositional strike	Concave sides of channel bends
Head of alluvial fans	Convex banks of channel bends
Points of exit of highland rivers onto alluvial plan	Heads of mid-channel bars
Regional unconformities	Point bars with suction eddies
Strand-line deposits	Scour holes, especially at tributary confluences
Incised channelways	Inner bedrock channels and false bedrock
Pediment mantles	Bedrock riffles
	Constricted channels between banks and bankward-migrating bars
Small Scale (10^0 m)	
Scoured bases of trough cross-strata sets	
Winnowed tops of gravel bars	
Thin ripple-form accumulations	
Dune Crests	
Dune Foresets	
Plane parallel laminae	
Leeward side of obstacles	

Note: See Slingerland and Smith (1986) for references concerning accumulation sites. After Slingerland and Smith (1986)

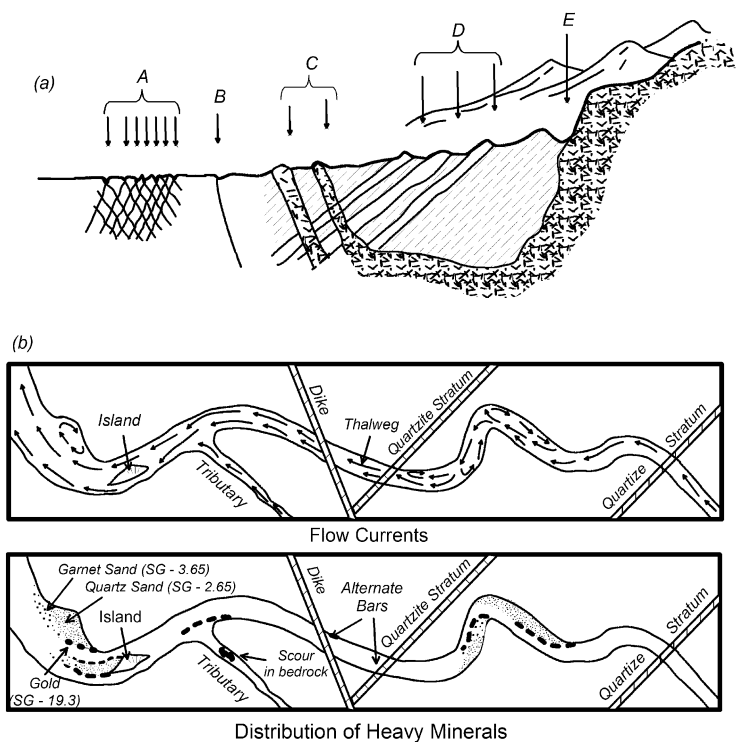


Figure 5.22. (A) Favorable sites for placer formation along a bedrock channel; (B) possible sites of heavy particle accumulation along alluvial channels (Modified from Guilbert and Park 1986)

allow for the efficient removal of hydraulically lighter materials from the channel bed sediment. For example, they examined the accumulation of magnetite within three sluiceways formed between a stable bank and a migrating sand bar in two rivers in South Africa. The zones were characterized by bar-top and sluiceway currents that are oriented across and downstream to the major flow direction, respectively. They found that at two sites magnetite concentrations were enriched over other parts of the river by approximately a factor of five. At both sites, bar growth and migration toward the bank had ceased because sediment swept into the sluiceway was immediately removed by the currents, leaving only the heavy mineral fraction. In contrast, heavy mineral concentrations were lacking at the third site where the bar was migrating toward the bank. In this case, the bar transported sediment was deposited in marginal foresets prior to being sorted by the sluiceway currents. The study by Smith and Beukes (1983) not only illustrates the need for the repeated scouring action associated with convergent flow, but demonstrates that even when heavy minerals are abundant, contaminant placers may not always develop.

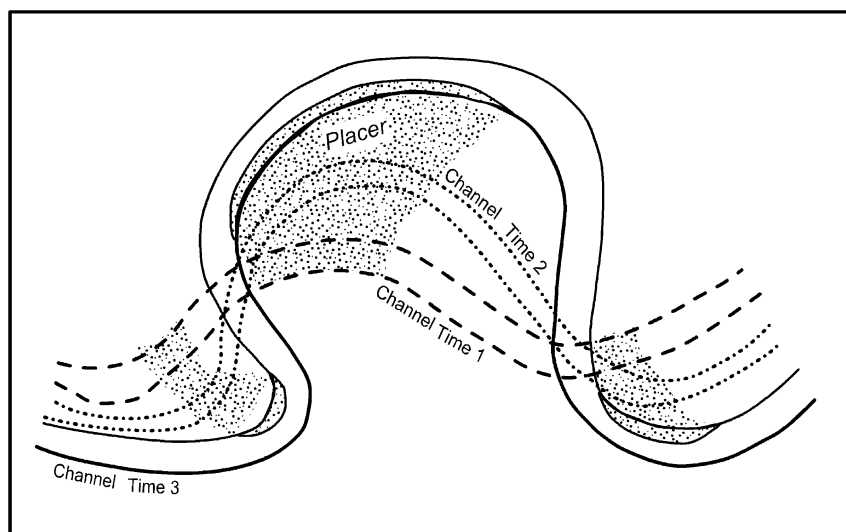


Figure 5.23. Formation of pay streaks (hotspots) in a laterally migrating meandering stream. Numbers represent different periods in time (Modified from Bateman 1950)

5.5.2.3. Variations dependent on time and frequency of deposition

Both the timing of deposition, and the frequency of inundation, can lead to differences in metal concentrations between morphological units. Graf et al. (1991), for example, found that within Queens Creek, a dryland river in Arizona, the active channel deposits inundated a few times per year possessed higher Cu, Zn, V, Mg, Mn, and Ti concentrations than those inundated about once per decade (Fig. 5.24). Similar observations have been made along the Rio Pilcomayo in southern Bolivia. In this case, trace metal concentrations observed in high-water channel deposits which are inundated during the wet season were low in comparison to low-water channel deposits that are inundated throughout the year, in spite of the fact that the low-water deposits are coarser grained (Fig. 5.25) (Hudson-Edwards et al. 2001).

The influence of inundation on metal concentrations is primarily related to two factors. First, the more frequent inundation of the lower, in-channel, morphological units provides for a greater opportunity to accumulate contaminated sediment, particularly fine-grained particles which are most likely to be transported during small to moderate floods (Graf et al. 1991). Second, deposition on topographically higher morphological units occurs only during major floods when significant quantities of sediment are delivered to the river from the entire drainage basin. The addition of relatively clean sediment from uncontaminated tributaries acts as a dilutant which decreases the metal concentrations within deposits formed during high flow conditions (Graf et al. 1991; Hudson-Edwards et al. 2001). During low flow, tributaries are unlikely to deliver water or sediment to the channel in significant

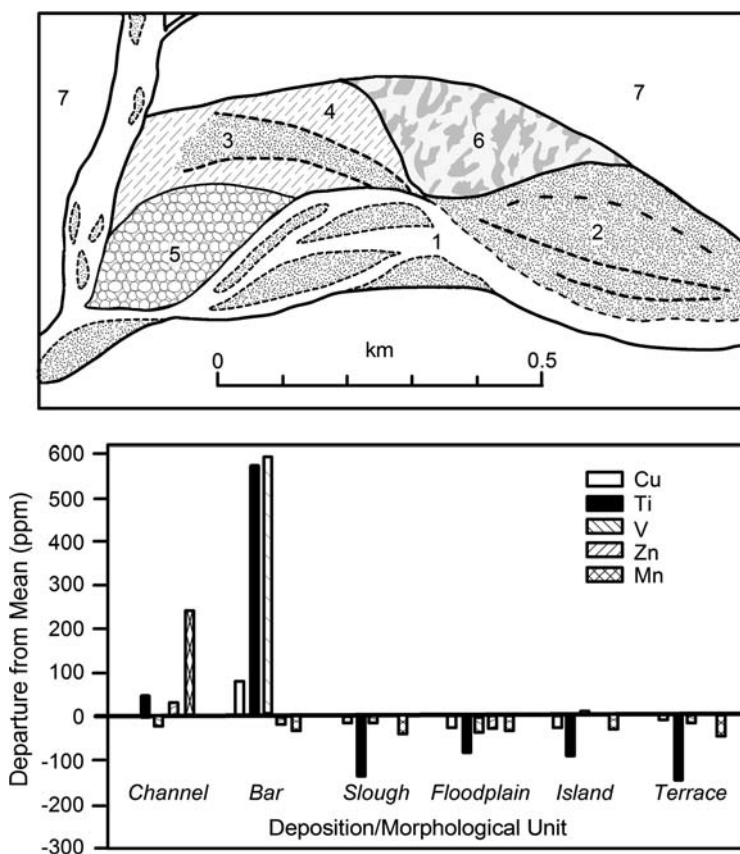


Figure 5.24. (A) Depositional environments along Queens Creek, Arizona, including: (1) active channel, (2) active bar, (3) slough, (4) floodplain, (5) island, (6) terrace, and (7) hillslopes; (B) Metal concentrations in selected depositional environments near Hewitt Canyon (From Graf et al. 1991)

quantities. Sediment-borne trace metals from pollutant sources, however, may enter the river year round with the discharge of waste waters. In the case of the Rio Pilcomayo, mentioned above, contaminated effluent from upstream ore processing facilities served as the primary source of water to the channel during the dry season.

5.5.2.4. Geochemically dependent variations

Differences in trace metal concentration between morphological units have also been attributed to localized variations in the deposits' physiochemical conditions, such as their ability to accumulate reactive coatings. A common example is the formation of Fe and Mn oxides and hydroxides on the surface of large immobile clasts. Once formed, the coatings may scavenge and accumulate metals from the water column (Chao and Theobald 1976; Tessier et al. 1982; Ladd et al. 1998). Moreover, scavenging by the coatings may be more significant within the channel bed deposits

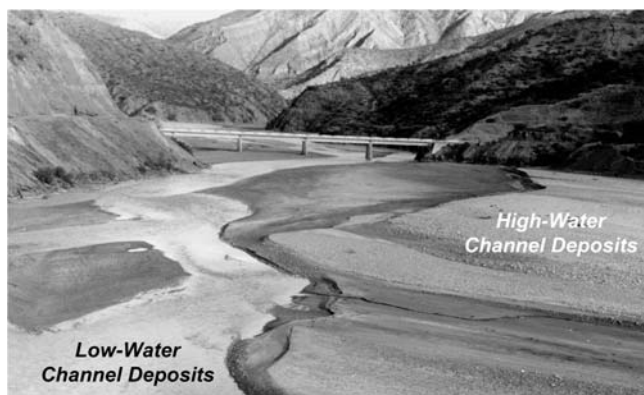


Figure 5.25. Rio Pilcomayo, southern Bolivia near Uyuni. Photo taken in July during the dry season

than on the adjacent, topographically higher bars which are finer-grained and less frequently inundated. In other cases, differences in the inundation and exposure of morphological units can lead to fluctuations in geochemical processes that differ between morphological units through time. These geochemical variations can lead to distinct differences in metal speciation and their potential for remobilization. The Rio Pilcomayo of southern Bolivia serves as an excellent example. Sulfide minerals deposited in high-water channel deposits (Fig. 5.25) are oxidized during the dry season creating localized zones of acid drainage enriched in trace metals (Miller et al. 2002). Presumably, the acid waters are flushed from the high-water deposits during the onset of the wet season. In contrast, sulfide oxidation within the adjacent low-water channel deposits which are continuously inundated is thought to be minimal, allowing for their downstream transport with only minor alteration. Thus, the metals remain locked within the sulfide grains.

5.5.3. Implications to Sampling

Collecting and analyzing multiple samples from a short-section of stream channel would reveal that metal concentrations within those samples vary from site to site. Birch et al. (2001) has referred to these local variations as *small scale* or *field variance*. The magnitude of field variance can be determined by the analysis of replicate samples from a given site; thus, it includes variations associated with the analytical methods. While analytical variation (error) is likely to be on the order of 5%, field variance is commonly on the order of 10 to 25% relative standard deviation for elements such as Cu, Pb, and Zn within fluvial systems (Birch et al. 2001).

As we previously noted, the problem with field variance is that it can significantly hinder many types of analyses, such as the assessment of large-scale, spatial trends in contaminant levels used to identify pollutant sources or downstream dispersal rates. In other words, where field variation is substantial, the ability to decipher

differences in contaminant levels between sample sites is considerably reduced, especially where concentrations approach background values (Birch et al. 2001). Thus, it is necessary to devise a sampling methodology that reduces field variance in order to document spatial trends over a broader area. While the need for such a program is widely understood, actually creating one is much more difficult. In fact, it has been argued that more money, effort, and time have been wasted because of poor sampling design than for any other reason (Keith et al. 1983; Horowitz 1991). The result has been the publication of numerous books on the topic (e.g., Watterson and Theobald 1979; Sanders et al. 1983; Keith 1988). Nonetheless, there is still no widely accepted strategy to the design of sediment sampling programs. Traditionally, many program managers have advocated a statistical approach in which samples are collected at a set distance (e.g., every 10 m across the channel), or on the basis of some form of gridded sampling design. The advantage of these methods is that they readily lend themselves to statistical manipulations (Provost 1984; Gilbert 1987; Mudroch and MacKnight 1991; Birch et al. 2001). However, our previous discussions show that trace metals are not likely to be randomly distributed within the channel bed sediment, but vary as a function of the sedimentology and associated hydraulic regime of the morphological units. In order to thoroughly characterize the concentrations that exist, each of the morphological units should be sampled. Performing this task using a random sampling approach would necessarily require a large number of samples to avoid missing some of the morphological units. The potential inefficiencies of the random sampling approach prompted Ladd et al. (1998) to suggest that investigators should initially conduct reconnaissance level surveys to determine the degree of variability that exists between morphological units before completing full-scale studies of metal distribution and their environmental impacts. If differences in concentration exist, they argue that the sampling program should stratify the collected samples (data) by morphological unit type.

Depending on the morphological complexity of the channel bed, it may be financially unrealistic to sample and analyze sediments from each of the morphological units at every sampling site. Thus, for some types of analysis, particularly those attempting to document spatial variations in metal concentrations, the sampling of only one morphological unit is advocated. Many exploration geologists use this approach when attempting to identify the location of potential ore bodies on the basis of spatial variations in the chemistry of the channel bed sediments (e.g., Day and Fletcher 1989). It should be recognized, however, that such an approach will not provide a full range of concentrations that exist (Ladd et al. 1998). Nonetheless, where this is necessary an understanding of the relationships between metal concentration and morphological unit may still lead to a more efficient sampling program than would be otherwise possible. Ladd et al. (1998) point out, for instance, that the full range of concentrations could be determined by sampling only two morphological units, those that possess the maximum and minimum concentrations within the river.

In addition to the variations in concentration that occur between morphological units, variance also exists within any given unit. This variance tends to increase

with the sedimentological and morphological complexity of the deposits (Birch et al. 2001). In order to reduce the variance inherent within the morphological units, composite sampling is commonly utilized in which multiple samples from a given morphological feature are collected, combined, and mixed prior to analysis. The primary advantage of composite sampling is that it reduces small scale sampling variance while decreasing analytical costs. It is once again important to recognize, however, that the generated results represent an average concentration of the contaminants within the morphological unit, rather than the full range of concentrations that exist.

5.6. PHYSICAL AND MATHEMATICAL MANIPULATIONS

In the preceding section it was shown how small scale variations in sediment-borne trace metal concentrations could be minimized by composite sampling of specific morphological units. In some cases, however, large scale variations in sediment size or composition cannot be removed by stratifying the data according to depositional environment. For example, a commonly observed trend along many rivers is for grain size of the channel bed material to decrease downstream. While sampling specific morphological units may adequately remove the local (reach scale) variations in grain size, and thus, metal concentrations, the larger scale, downstream variations will remain. These types of variations in sediment size and/or composition may reduce our ability to decipher spatial (or temporal) patterns in sediment geochemistry. It is therefore common to apply some form of stratification to the collected samples or data to compensate for variations in grain size or composition. Two approaches are frequently utilized. First, a specific grain size or sediment type can be removed and analyzed independently. The separation and analysis of a specific grain size fraction is widely advocated for environmental studies. The assumption inherent in this procedure is that trace metals are contained entirely within the analyzed fraction. While the actual size of the chemically active sediment will vary from river to river, many investigators have argued that the grain size of the analyzed sediment should be standardized to allow for comparison of different investigations conducted in diverse environments. In the U.S., most analyses, including those conducted by state and federal regulatory agencies, are based on the collection and analysis of the $< 63 \mu\text{m}$ sediment fraction. The advantages of using the $< 63 \mu\text{m}$ fraction over other size ranges are that: (1) this size fraction can be extracted from the bulk sample relatively quickly via sieving, a process that does not alter trace metal chemistry, (2) the $< 63 \mu\text{m}$ particle size is most frequently carried in suspension in riverine systems, and may therefore be most readily distributed over the environment, and (3) it occurs in large enough quantities that it can generally be collected from both the water column and alluvial deposits (Miller and Lechler 1998). A significant disadvantage, however, is that the analyses do not provide for an understanding of the actual concentrations in the bulk sample which may be required for many risk assessments.

The second approach is to mathematically manipulate geochemical data obtained from the bulk sample using information collected from the analysis of a separate subsample of the analyzed material. The most common form of mathematical manipulation involves normalization of bulk concentrations. The first step in normalization is the calculation of a dilution factor describing the amount of a substance within the sample which is assumed to be devoid of trace metals; a normalized concentration can then be calculated by multiplying the dilution factor by the chemical concentration determined for the bulk sample. The following equations, for example, could be used to normalize data with respect to the proportion of sediment $< 63 \mu\text{m}$ in size:

$$(10) \quad \text{Dilution Factor} = 100 / (100 - \% \text{ of sample } > 63 \mu\text{m in size})$$

$$(11) \quad = 100 / (\% \text{ of sample } < 63 \mu\text{m in size});$$

$$(12) \quad \text{Normalized Concentration} = (\text{dilution factor}) \\ \times (\text{conc. metal in bulk sample})$$

In this case, sand-sized sediment is the diluting substance.

Although mentioned above, it is worth reiterating that calculated normalized data are used for clarifying spatial or temporal trends; they do not reflect the actual chemical concentrations within the alluvial sediments (Horowitz 1991). For example, Table 5.6 presents metal concentrations for the $< 63 \mu\text{m}$ sediment fraction determined by analyzing the fine materials separated from the bulk sample, and by normalizing the data according to the percentage of sediment $< 63 \mu\text{m}$ in size. Differences in the normalized and measured concentrations can be significant, especially when the samples contain $< 50\%$ silt and clay (Table 5.6). The observed differences may indicate that: (1) analytical errors are associated with the geochemical or grain size analyses, (2) not all of the sediment-borne trace metals are associated with the analyzed grain size fraction, and/or (3) chemical variations in trace metal concentrations are dependent on factors other than simply grain size (Horowitz 1991).

Normalization by grain size is by far the most frequently used form of data manipulation. Nonetheless, bulk concentrations have also been normalized according to the percentage of quartz, carbonate, or organic carbon within the sample (Table 5.7). For each substance, the assumption is that particles of a specific composition are free of substantial quantities of trace metals. Note, however, that organic carbon can act both as a dilutant and a concentrator of trace metals, depending on its form and the physiochemical conditions of the site.

A slightly different approach is to normalize bulk trace metal concentrations by the concentration of a conservative element such as Al, Ti, or Li. The underlying assumption is that the conservative elements are released from crustal rocks at a uniform rate when considered over a long enough timeframe. Thus, areas of trace metal concentration or dilution can be identified when normalized by the conservative element. In contrast to the methods used for grain size, quartz, etc., normalization in this case is performed by simply dividing the concentration of the

Table 5.6. Comparison of calculated and measured trace metal concentrations based on the < 63 µm sediment fraction; concentrations in mg/kg

Location	Percent < 63 µm	Data	Cu	Zn	Pb	Ni
Georges Bank (M8-5-4)	1	Measured	104	55	22	20
		Calculated	100	500	500	100
Columbia Slough	18	Measured	47	225	72	31
		Calculated	145	811	239	145
Nemadji River	44	Measured	31	62	10	31
		Calculated	50	102	20	48
Patuxent River at: Point Patience	54	Measured	35	116	39	32
		Calculated	43	180	46	41
Yaharra River	66	Measured	22	45	28	22
		Calculated	20	41	33	17
Hog Point	76	Measured	25	131	26	31
		Calculated	26	147	29	36
George Bank (M13A)	89	Measured	10	63	22	22
		Calculated	15	67	22	25
Lake Bruin	95	Measured	24	104	23	29
		Calculated	26	108	23	31

Modified from Horowitz (1991)

Measured – actual analytical determination of < 63 µm sediment fraction; Calculated – based on product of bulk chemical concentration and a dilution factor; DF = 100/100-% of sample > 63 µm in size

Table 5.7. Selected dilution factors (DF) and normalization equations used to manipulate bulk geochemical data

Equation	Comment
Grain Size DF = 100/(100-% Grain Size > Size of Interest)	Size is typically > 63 µm in US.
NC = (DF)(Bulk Chemical Concentration)	
Carbonate DF = 100/(100-% Carbonate)	Normalization to carbonate-free basis; used primarily in marine or karst environments
NC = (DF)(Bulk Chemical Concentration)	
Organic Carbon Content DF = 100/(100-% Organic Carbon)	Normalized to organic carbon-free basis; loss-on-ignition commonly used as estimate of % o. carbon; organic carbon does not serve as a dilutant in many systems, but can accumulate trace metals
NC = (DF)(Bulk Chemical Concentration)	
Conservative Element NR = Concentration of trace element / Concentration conservative element	Produces a ratio rather than a value in concentration units; commonly used conservative elements include Al, Li, and Ti.

Adapted from Horowitz (1991)

NC – Normalized concentration of trace metal

NR – Normalized ratio

trace metal by the concentration of the conservative element. The resulting values are not in units of concentration, but represent a ratio. The use of a ratio can make comparisons with data from other regions difficult (Horowitz 1991). As a result it is probably fair to say that normalization to a conservative element is not frequently used in environmental studies.

It is extremely important to recognize that normalization procedures should only be used when there is some quantitative rationale for doing so (e.g., a strong correlation between trace metal concentrations and percent silt and clay in the sample). Even in this case, care must be taken in that misleading results can be produced. For instance, normalization by grain size is commonly inappropriate for many mining impacted rivers because the trace metals are distributed across a broad range of particle sizes (Moore et al. 1989). Unfortunately, many regulatory agencies require sampling protocols to be based on only the fine-grained sediment fraction, or some form of mathematical normalization, without ever determining if it is appropriate for the site of interest. The results can lead to a misunderstanding of both the quantity and the distribution of trace metals in the system.

5.7. TEMPORAL VARIATIONS IN CONCENTRATION

Viganò et al. (2003) collected bed sediments along the Po River in Italy during two sampling campaigns: one during the summer, 1996, and the other during the winter of 1997. Comparison of the results revealed that trace metal concentrations at a given sampling site were similar and, therefore, largely independent of the season during which the samples were collected (Fig. 5.26). Their conclusions emphasize one of the commonly cited advantages of conducting geochemical surveys of channel bed sediment – that the data are largely devoid of the temporal variations in concentration observed for the dissolved or suspended load. Bed sediment, in other words, probably serves as an indicator of both river health and spatial variations in contaminant levels without requiring multiple sampling campaigns conducted on an event or seasonal basis. Although this assumption is founded on a surprisingly limited number of investigations, it is generally supported by the existing data, even though seasonal variations in trace metal concentrations in channel bed sediments have been identified along some rivers or river reaches. For example, Gaiero et al. (1997) found that in uncontaminated (pristine) areas of the Suquia River system of Argentina, differences in total non-residual trace metal concentrations between the spring and autumn were minimal. However, in contaminated areas affected by untreated sewage, trace metal concentrations were higher in the springtime than in the fall. The temporal variations were attributed to the type and abundance of organic matter and its effects via Eh on Mn and Fe hydroxide dissolution (see Chapter 2). The influence of organic matter cycling on seasonal variations in channel bed concentrations have also been suggested for other rivers (see, for example, Facetti et al. 1998). Nonetheless, it is important to recognize that while seasonal variations in concentrations at a specific site may occur for some constituents, the

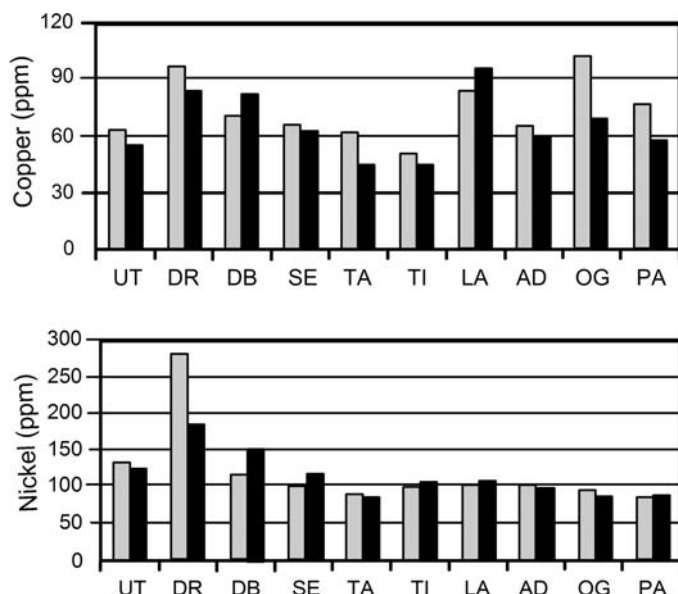


Figure 5.26. Concentrations of Cu and Ni in the $< 63 \mu\text{m}$ fraction of channel bed sediments from the Po River, Italy. Samples were collected in the summer (grey bars) and winter (black bars). Acronyms along x-axis represent successive downstream sampling sites. Note minimal variations in concentration between seasons (Modified from Viganò et al. 2003)

spatial (downstream) trends typically remain consistent during the year, provided that the character of the contaminant sources remain relatively unchanged.

A number of studies have examined the impact of high magnitude, low frequency floods on channel bed concentrations. For the Carson River, Nevada, Miller et al. (1999) found that neither the concentrations of Hg at a site, nor the downstream trends in Hg concentration in the channel bed were altered during a 100-year event in 1997 (Fig. 5.8b). The observed stability of the Hg concentrations was apparently controlled by the overall geomorphic structure of the system, defined by its valley morphology, the location of tributaries that delivered “clean” sediment to the channel, and the distribution of Hg within the valley fill.

The alteration and rapid recovery of channel bed concentrations following a 100-year event was observed by Ciszewski (2001) along Biala Przemsza River of southern Poland. Here trace metal concentrations decreased by three fold over a 40 km reach of the river in response to the flood, but during the following period (August 1997–March 1998) concentrations recovered to values observed in 1993 prior to the event. A second, more moderate flood approximating the bankfull discharge, resulted in inconsistent downstream changes in channel bed concentrations; downstream reaches exhibited decreases in concentration, whereas upstream reaches lying adjacent to the influx of waste waters from a mine exhibited increases in concentration. The Biala Przemsza study demonstrates that concentrations of

trace metals in the channel bed can be altered by flood events, but may rapidly return to a mean condition.

While flood induced variations in trace metal concentrations in the channel bed appear to be limited, recognize that the depth of scour during a major flood can be significantly greater than at other times. The eroded materials can then be redistributed over the valley floor, thereby changing the spatial distribution of the sediment-bound contaminants observed prior to the event. The extent of metal redistribution can be particularly important in areas where significant quantities of trace metals have been stored beneath the channel as a result of aggradation. In this case, erosion of older, highly contaminated sediments can actually increase the concentrations observed at the surface of the adjacent floodplain and re-contaminate previously remediated areas (NRC 2005).

5.8. SUMMARY

The use of channel bed materials to determine ecosystem health, contaminant sources, and sediment-borne transport dynamics greatly simplifies site assessments as bed materials are easier to collect, can be obtained at most times throughout the year, and exhibit less variations in concentration through time than does suspended sediment. The ability of a river to entrain, transport, and deposit sediment and sediment-borne trace metals depends on the energy given to the water by velocity, depth, and slope and on the amount of energy consumed by the resistance to flow dictated by such elements as channel configuration, particle size, and sediment concentration. Contaminant sources and dispersal processes combine to produce longitudinal trends in trace metal concentrations that are river specific. Nonetheless, concentrations tend to decay quasi-systematically downstream of point sources. Observed decay patterns are dependent on the local importance of distinct, but interrelated dispersal mechanisms including: (1) hydraulic sorting, (2) sediment storage and exchange with the floodplain, (3) dilution associated with the mixing of contaminated and uncontaminated sediment, (4) biological uptake, and (5) geochemical remobilization or abstractions. Within many rivers, sediment transport and deposition is not uniformly distributed downstream. Rather, one or the other process is locally dominant creating alternating zones of transport and sedimentation that affect the general longitudinal trends in trace metal concentration. Sediment-borne trace metals may also move through the system as a series of sediment slugs, complicating geographical patterns predicted by regression or mixing models.

The in-channel storage of trace metals within stable channels is generally small, accounting for less than about 10% of the total load exported from the basin. This is in marked contrast to the storage of sediment within floodplains, where more than 40–50% of the total annual load may be deposited. The residence time of the contaminants also differs substantially, ranging from less than 5 years to decades, centuries, or millennia for channels and floodplains, respectively. At the reach scale, sediment-borne trace metals are selectively deposited within distinct morphological

units (e.g., bars, riffles, pools, and chutes) as the result of hydraulic sorting by size and density, the timing and frequency of inundation, and various geochemical processes. These morphologic features are both created by and influence the mean flow configuration through the reach, and, therefore, have specific associations with a given channel pattern (e.g., point bars, pools, and riffles with meandering streams, braid bars with braided channels, etc.). Channel patterns can be broadly classified as straight, meandering, braided, and anabranching forms. However, the boundaries between the different patterns are indistinct and may change both along a given river and as a function of discharge at a specific site.

While protocols aimed at sampling specific morphological units may adequately remove local (field scale) variance in trace metal concentrations, larger scale (downstream) variations may remain as a result of differences in particle size or composition. The net result is that our ability to decipher spatial (or temporal) patterns in sediment geochemistry is greatly reduced. It is therefore common to apply some form of stratification to the collected samples or data to compensate for variations in grain size and/or composition. Two approaches are frequently used; a specific grain size or sediment type can be removed and analyzed independently, or bulk sample concentrations can be mathematically manipulated using data from a separate sample split to remove the effects of size and/or composition. The approach used necessarily depends on the sediment-trace metal relations within the channel bed sediment.

5.9. SUGGESTED READINGS

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Miller, J.R.; Orbock Miller, S.M.

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