Section 2 – Channel Mechanics

Introduction

- Near the end of the previous section, we recognized that direct precipitation into a stream channel is responsible for the initial flood response. Not only is the response of the stream to this input the fastest, it is also the largest in terms of instantaneous magnitude or discharge.
- We will discuss stream discharge in more detail a little later. The aim here is to investigate the flow of water in an open channel. How does the water interact with itself and the channel through which it is moving? What are some of the mathematical relationships used to quantify open channel flow?
- Although the processes by which water moves over and through drainage basin slopes are geomorphologically important, it is water flowing within channels that perhaps does the most geomorphological work.
- By carving the channels in which they flow, rivers create most of the relief available for slope processes to act upon. They also transport debris generated by these processes to ocean basins. And they produce a suite of erosional and depositional landforms.
- The initiation of a small rill or gully by a trickle of water leads to the creation of a complex drainage pattern and development of large, continent-wide drainage basins. Geomorphology has moved beyond simple qualitative descriptions of the landscapes created by fluvial action, to a new era of quantitative geomorphology.
- Our ability to quantify the multitude of variables and mathematical relationships defining open channel flow is growing but we are still not able to apply this knowledge in a predictive sense. Much more data are required before we can truly model the full complexity of a fluvial system.
- The mechanics of fluid flow are much more difficult to understand than the static relationships presented by mass movements or slope processes because fluids are in constant motion (Newtonian viscous fluid).

- This branch of geomorphology has practical applications such as the restoration of aquatic habitats, remediation of contaminated rivers, and the design of flood control structures so it is not for the sole purpose of a theoretical understanding of river flow.

2.1 – Basic Mechanics of Open Channel Flow

- In order to understand stream flow, we need to understand the shape of the channel through which the water is moving. Some of the common descriptors of channel shape shown in figure 2-1 are:

  w – width of water in the channel,

  P – wetted perimeter given as $2d + w$ (i.e. the boundary along which water is in contact with the channel floor and walls),

  A – the area of transverse section of the stream,

  d – water depth

  R – hydraulic radius given as $A / (2d + W)$

  s – stream gradient or the drop in elevation between two points on the stream surface.

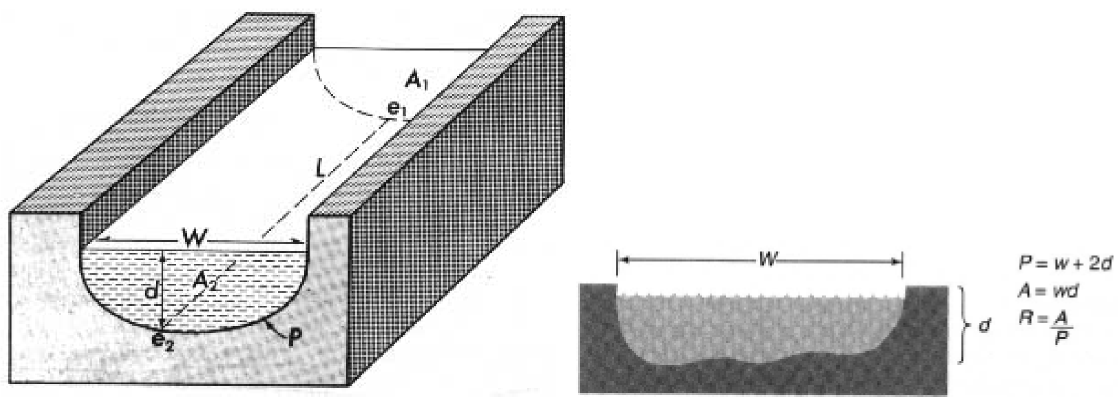

Figure 2-1. Stream channel dimensions (Selby 1991 and Ritter *et al*. 2002).

- Discharge of water through a smooth channel (Q) is the product of average velocity (v) and cross-sectional area (A) under steady uniform flow (i.e. $Q = Av$).
- A river is continuously working towards a state of quasi-equilibrium with the environment within which it is flowing. Rivers attempt to balance driving and resisting

forces according to how much potential kinetic energy is available to them and how much of that energy is consumed by friction, or resistance to flow.

- Because gravity is the main force acting on the water, under ideal conditions, the force of gravity would continuously accelerate the water in a downstream direction. In actual stream channels, this acceleration is balanced through an increase in turbulence resulting from increased velocity due to resisting components.
- To a large extent, it is the friction within the water (i.e. between water molecules) and between the water and stream channel (water molecules with solid particles) that defines the flow characteristics of a stream.
- Stream flow has been categorized into several different types according to flow character (Table 2-1).

Table 2-1. Flow characteristics in open channels (Ritter *et al*., 2002).

Type of flow	Flow character
Spatial variations in velocity	
Uniform flow	Velocity is constant along the channel
Nonuniform (varied)	Velocity changes with distance along the channel
Temporal variations in velocity	
Steady flow	Velocity does not change in magnitude or direction with time
Unsteady flow	Velocity fluctuates in magnitude or direction with time
Degree of particle mixing	
Laminar flow	Fluid elements move along specific paths with no significant mixing among the adjacent layers; Re < 500
Turbulent flow	Fluid elements do not flow along parallel paths, but repeatedly move between adjacent layers; involve large-scale transfer of momentum across layer boundaries; Re > 2000

- Most streams exhibit nonuniform, unsteady, and turbulent flow along their entire length. The other flow types expressed in table 2-1 represent theoretical open channel flow under ideal conditions (i.e. no friction).
- Laminar flow may occur in some quiet reaches of a stream where there is no transfer of energy between layers of flow however turbulence is usually present to some degree.

Turbulence due to interaction of the water with the channel bed and banks causes much of the energy to be dissipated.

- Under conditions of laminar flow, energy is dissipated between layers flowing at different velocities due to hydraulic shearing.
- A dimensionless parameter called the Reynolds number (Re) can be used to estimate the point at which laminar flow changes to turbulent flow as depth and velocity increase.

$$Re = vR?/\mu$$

where *?* the density, and *µ* the molecular viscosity of water. Density and viscosity are not constants because they change in response to changes in temperature and/or sediment concentration.

- The Reynolds number, in fact, represents the relationship between driving forces (i.e. gravity expressed through velocity and depth) and resisting forces (the internal resistance of water molecules). If we consider that the ratio ?/µ represents a fluid property called *kinematic viscosity* (?) the equation can be rewritten as

$$Re = vR/?$$
$$= \text{driving forces/resisting forces}$$

- Based on experiments in the field and in flumes laminar flow occurs when Re $\leq$ 500 and true turbulent flow occurs when Re $\geq$ 1500. A further distinction can be made between tranquil turbulent flow, cricital turbulent flow, and rapid turbulent flow using the Froude number (Fr)

$$Fr = v\ /\ (dg)^{0.5}$$

where *g* is gravity. Tranquil turbulent flow occurs when $Fr < 1$, cricital flow is when $Fr = 1$, and rapid turbulent flow occurs under conditions where $Fr \geq 1$. The main difference between these turbulent flow types is the amount of energy that is dissipated.

- A relationship between type of turbulent flow and sedimentary bedforms exists as illustrated in figure 2-2 and 2-3.

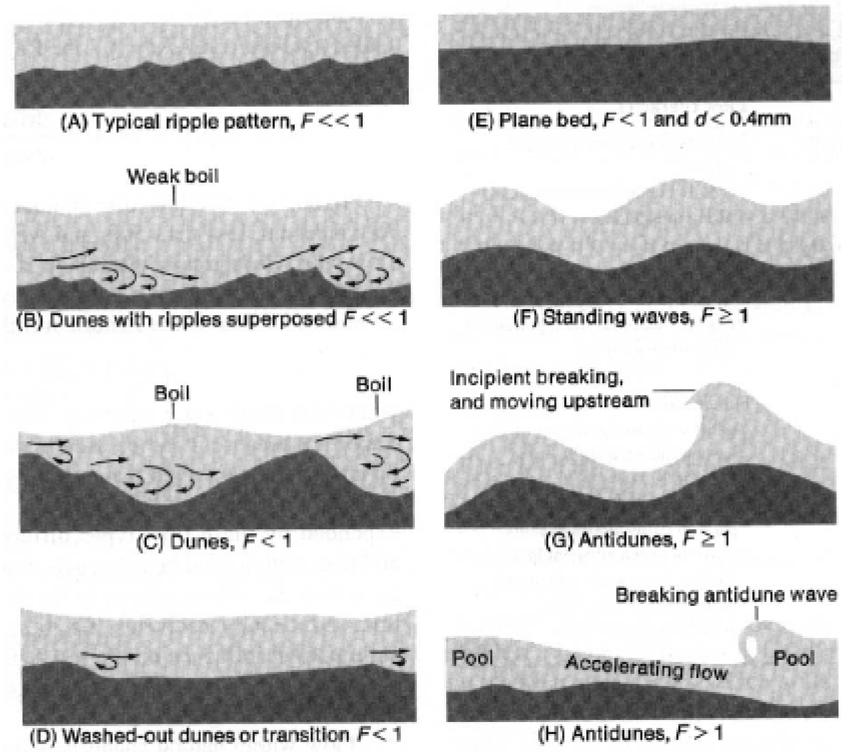

Figure 2-2. Bed forms in alluvial channels and their relation to flow conditions. *F* = Froude number, *d* = depth (Simmons and Richardson 1963 *in* Ritter *et al*. 2002).

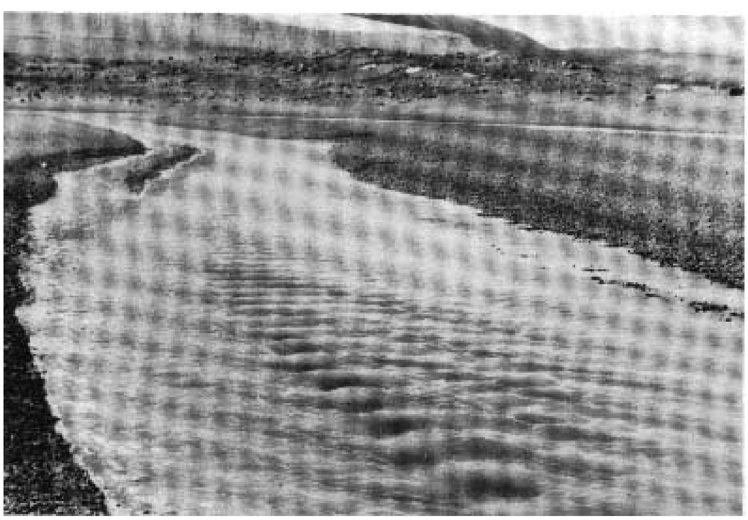

Figure 2-3. Standing waves above antidunes of sandy gravels in a melt-water stream, Antarctica (Selby, 1991).

- Geomorphologically, this is an important connection between bedforms that may have been preserved in a sedimentary record and our ability to interpret some aspects of past channel conditions (Fig. 2-4).

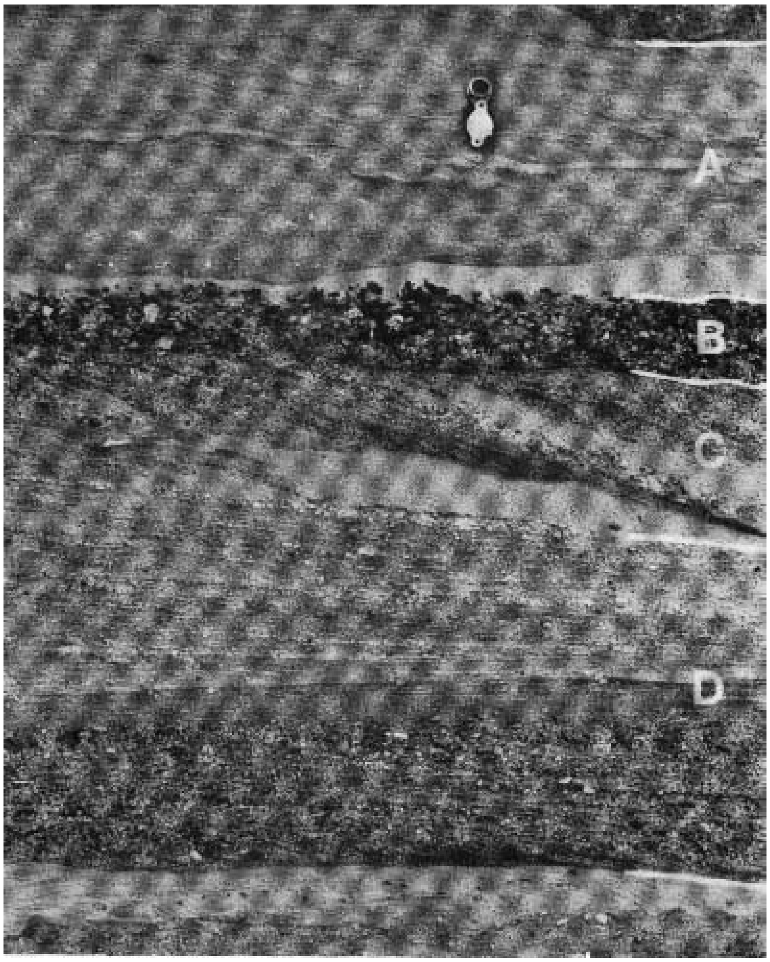

Figure 2-4. A, ripple bedding in sand; B coarse bedload with no bedform structure; C cross-bedding of the advancing faces of dunes; D planar bedding. Scale is given by the hand lens (photo by C.S. Nelson *in* Selby 1991).

- Most of the turbulence within a stream channel is generated at the interface between water and sediment. Thus as resistance increases, velocity decreases towards the channel perimeter (i.e. banks and bed). Figure 2-3 graphically demonstrates the relationship between channel velocity, depth and width.

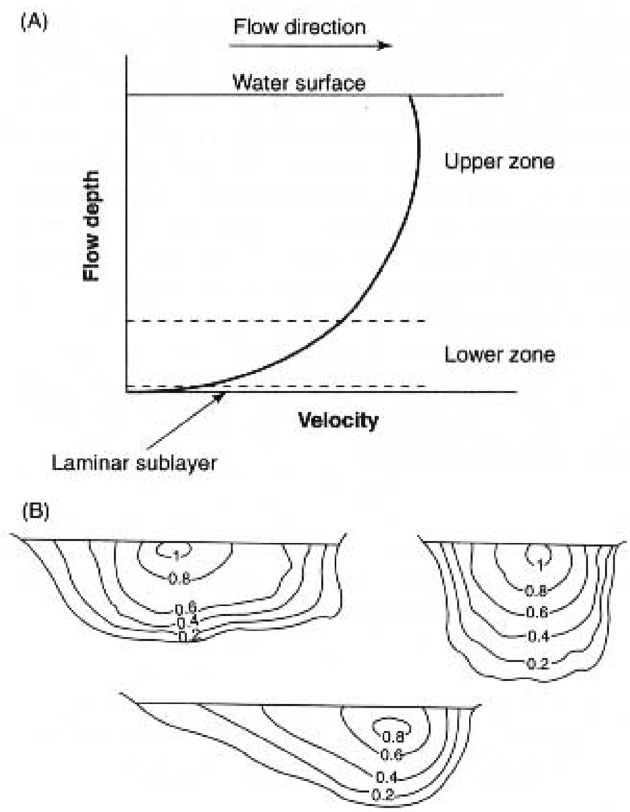

Figure 2-5. (A) Variations in flow velocity as a function of water depth. (B) Typical variations in flow velocity as a function of width. Isovels (lines of equal velocity) are in m/s. (Modified from Wolman 1955 *in* Ritter *et al*. 2002).

- As can be seen in figure 2-5, areas of highest velocity are located near the center of the channel and/or where depth is greatest.

- In channels of sufficient depth and formed in sand or finer-grained sediments with smooth channel beds, the velocity profile can be partitioned into an upper and lower zone (Fig. 2-5 (A)). The lower zone is heavily affected by bedform configuration and any associated resistance while the upper zone may be affected by wind action, especially at the water surface.

- Velocities are typically measured at 0.6 of the depth from the water surface, as this zone approximates the mean flow velocity in the channel. In rivers where the flow depth is very small compared to the size of roughness elements (i.e. bed material, boulders, clasts) the entire velocity profile closely resembles a parabolic curve.

2.2 – Velocity and Resisting Forces

- As has been demonstrated, the velocity of water in a stream channel is not uniformly distributed throughout the channel, nor is it independent of channel shape in terms of width, depth or cross-sectional area.

- Several mathematical relationships have been developed which compute flow velocity using hydraulic radius, slope (i.e. gravity) and various constants in order to account for the presence of resisting factors.

- The Chezy and Manning equations are frequently used to compute flow velocities which more closely correspond to those measured in the field because both equations incorporate a resisting factor which checks flow velocity.

- The Manning equation is the more popular method of estimating actual flow velocities though the Chezy and Manning equations are mathematically related. Velocity is calculated as

$$v = n\, R^{2/3} s^{1/2}$$

where *n* is the Manning roughness coefficient, a resisting element typically derived from a table or figure such as those shown below.

Table 2-2. Manning roughness coefficients (*n*) for different boundary types.

Boundary	Manning n ($ft^{1/6}$)
Very smooth surfaces such as glass, plastic, or brass	0.010
Very smooth concrete and planed timber	0.011
Smooth concrete	0.012
Ordinary concrete lining	0.013
Good wood	0.014
Vitrified clay	0.015
Shot concrete, untroweled, and earth channels in best condition	0.017
Straight unlined earth canals in good condition	0.020
Rivers and earth canals in fair condition; some growth	0.025
Winding natural streams and canals in poor condition; considerable moss growth	0.035
Mountain streams with rocky beds and rivers with variable sections and some vegetation along banks	0.041–0.050

(a) Floodplains	
short grass pasture	0.025–0.035
mature crops	0.025–0.045
brushwood	0.035–0.070
forested	0.050–0.160
(b) Alluvial channels	
smooth sand beds, no vegetation	0.014–0.035
dunes on channel beds	0.018–0.035
smooth beds with pools and water weeds	0.045–0.080
(c) Mountain streams	
gravel and few boulders	0.030–0.050
large boulders	0.040–0.070

Note: Values given are for straight channels. For winding channels these values should be increased by up to 30 per cent, depending on the degree of winding.
Source: Chow, 1959.

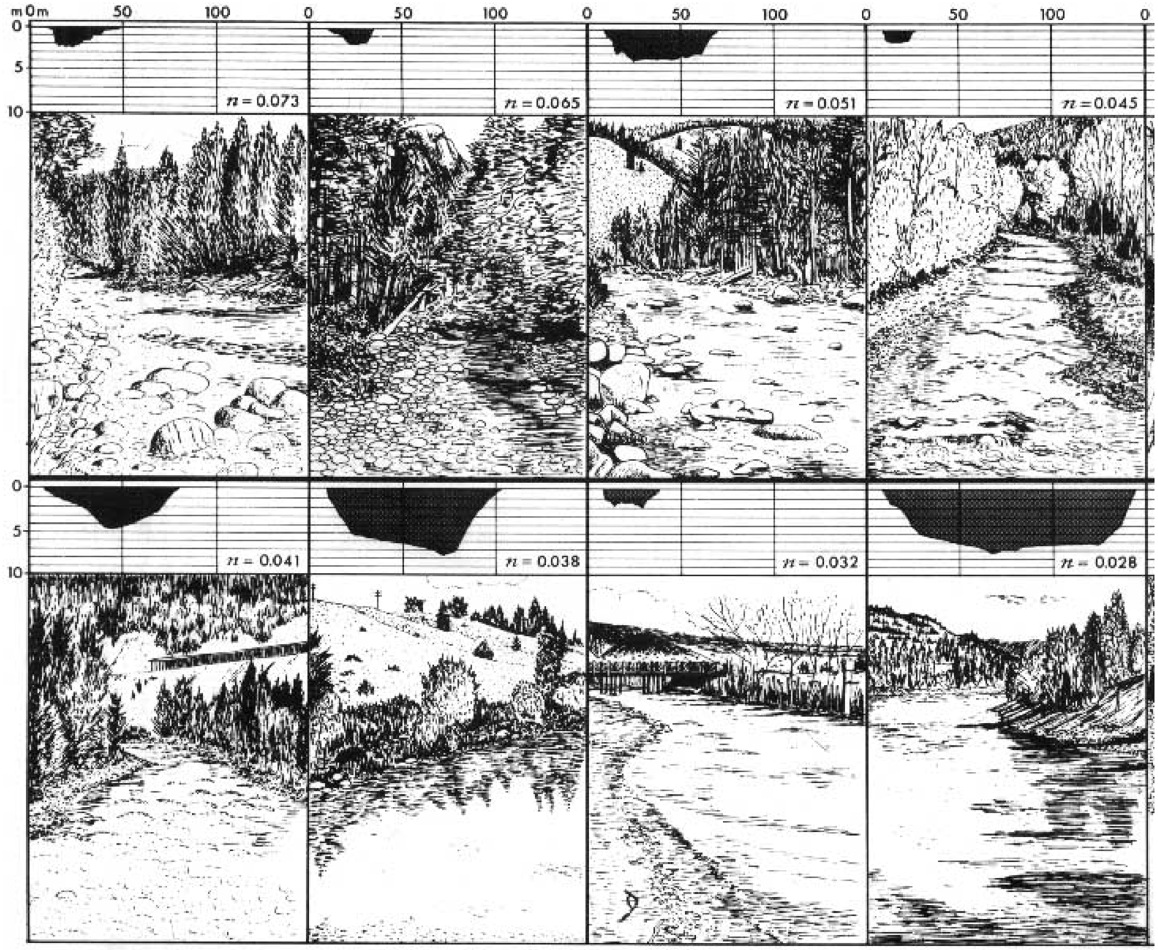

Figure 2-6. Representations of various stream channels with cross-sections and values of Manning's *n* (based on photographs in Barnes 1967 *in* Selby 1991).

- Although Manning's *n* is assumed to be constant for any particular channel framework, in reality, it varies with flow stage because channels become more efficient as discharge increases.

- All resistance coefficients are estimated based on hydraulic variables (eg. s, R, v) because they cannot be measured directly. Furthermore, they are dependent on many other factors which are normally not taken into consideration such as particle size, sediment concentration, and bottom configuration.

- There must be a trade off between accuracy and function. If, in order to estimate Manning's *n* we had to include all of the variables mentioned above, we'd need to have detailed knowledge about a channel or stream system where in fact what we're trying to do is estimate its cricital parameters using proxies.

- Although Manning's n represents the total resistance within a channel, resistance can be subdivided into three main components, namely free surface, channel, and boundary.

- Free surface resistance is the loss of energy from disruption of the water surface by surface waves caused by wind or hydraulic jumps (eg. rapids). Channel resistance is associated with the configuration of the bed and cross-sectional area. Boundary resistance is caused by individual clasts or microtopographic features of the channel.

- Microtopographic features produce form drag which is a function of depth and grain size or roughness. A measure of relative roughness can be computed through the ratio between water depth (d) and dominant grain size in the channel (D_i) or d / D_i.

- Following the work of Wolman (1955), an equation was developed which expresses the resistance-particle-size relationship in the form of the Darcy-Wesibach formula states as

$$F = (1 / 2\log d / D_{84} + 1)^2$$

where F is the Darcy-Weisbach resistance parameter and D_{84} is the particle diameter that is equal to or larger than 84 percent of the clasts on the channel bottom.

- The above expression implies that flow resistance decreases with increasing depth or decreasing particle size. Although D_{84} is rather arbitrary, it is consistent with the fact that the largest particles in the bed play the dominant role in controlling flow resistance.
- Rivers rework the channel bed and bedload in order to maximize efficiency and minimize resistance within the limits of their capacity to move the material being deposited into the channel.
- For example, there is little that small streams can do to rearrange thier bed configuration following the deposition of material from a rock fall event into the channel

(Fig. 2-7). In most cases, however, the stream has sufficient power to redistribute at least some of the sediment.

Figure 2-7. Boulders with dimensions of 10m partly block the channel of the Haast River. The largest debris which can be rolled along the bed has dimensions of about 3m, but travel distances are short because of channel roughness (Selby 1991).

- Rivers in which the material being transported is relatively fine grained (i.e. much smaller than gravel) will experience resistance not only from individual particles, but more importantly from the larger-scale configuration of bedform features.
- Other factors affecting flow resistance include channel bars, vegetation along banks which protrudes into the river, and sediment load or concentration. The precise contribution of vegetation to flow resistance is difficult to quantify because of the diverse nature of it.
- The relationship between sediment concentration on Manning's *n* was first described by Vanoni (1941, 1946) as shown in figure 2-8.

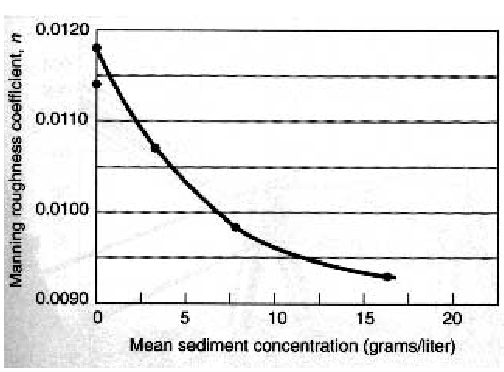

Figure 2-8. Effect of suspended load concentration on the Manning's *n* (Selby 1991)

- Increased suspended load appears to dampen mixing which in turn reduces turbulence and thus diminishing resistance. Also, the velocity of highly concentrated stream flow should be higher than clear water flow due to decreased turbulence and increased gravitational force.