# Watershed Fundamentals

## INTRODUCTION

Making decisions about a watershed is an important responsibility; decisions must be based on a solid understanding of the characteristics of the watershed and how physical processes shape watershed conditions. This section provides basic background information on watershed functions and processes to help users understand the assessment procedure and the results of the assessment process. Watershed "processes" refer to those natural physical, chemical, and biological mechanisms that interact to form aquatic ecosystems. For example, the input and routing of water, ***sediments***,[1] and large wood through stream channels involve many inter-related processes occurring both in-channel and upslope.

## WHAT IS A WATERSHED?

The term "watershed" describes an area of land that drains downslope to the lowest point (Figure 1). The water moves by means of a network of drainage pathways that may be underground or on the surface. Generally, these pathways converge into a stream and river system that becomes progressively larger as the water moves downstream. However, in some arid regions, the water drains to a central depression such as a lake or marsh with no surface-water exit.

Watersheds can be large or small. Every stream, tributary, or river has an associated watershed, and small watersheds aggregate together to become larger watersheds. It is a relatively easy task to delineate watershed boundaries using a topographical map that shows stream channels. The watershed boundaries will follow the major ridge-line around the channels and meet at the bottom where the water flows out of the watershed, commonly referred to as the mouth of the stream or river.


**Figure 1. Watershed is an area of land that drains downslope to the lowest point.**

The ***connectivity*** of the stream system is the primary reason why aquatic assessments need to be done at the watershed level. Connectivity refers to the physical connection between tributaries and the river, between surface water and groundwater, and between wetlands and these water sources. Because the water moves downstream in a watershed, any activity that affects the water quality, quantity, or rate of

[1] Terms found in bold italic throughout the text are defined in the Glossary at the end of this component.



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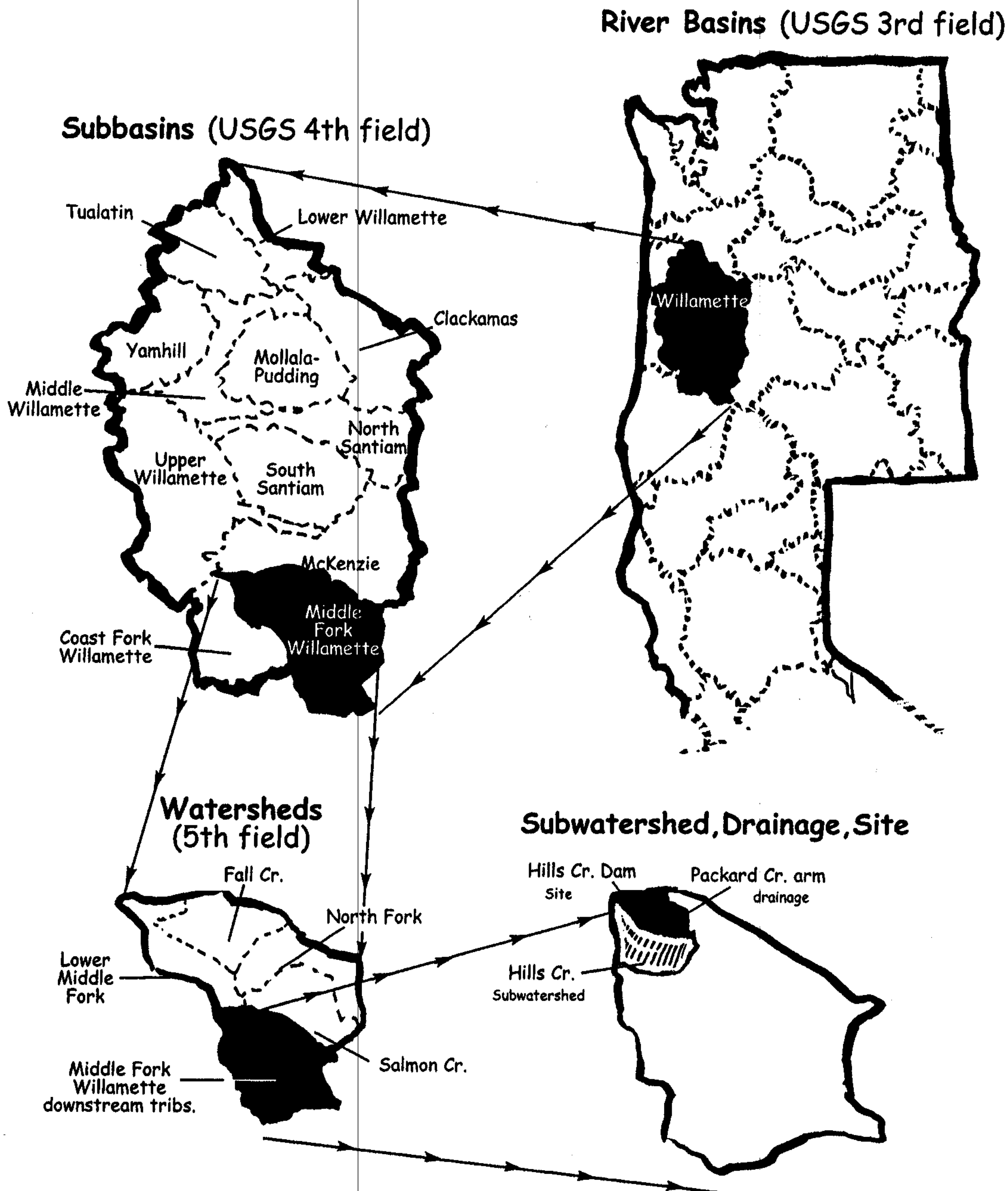


**Figure 2. Suggested terminology for watershed descriptive terms based on USGS hydrologic "fields." These fields correspond to the following terms: river basin (3rd field), sub-basin (4th field), and watershed (5th field). In the figure, the Willamette River Basin is divided into sub-basins including the Middle Fork Willamette, which is divided into watersheds including the Middle Fork Willamette downstream tributaries. This watershed then includes a subwatershed, drainage, and site, as seen in the lower right of the figure.**

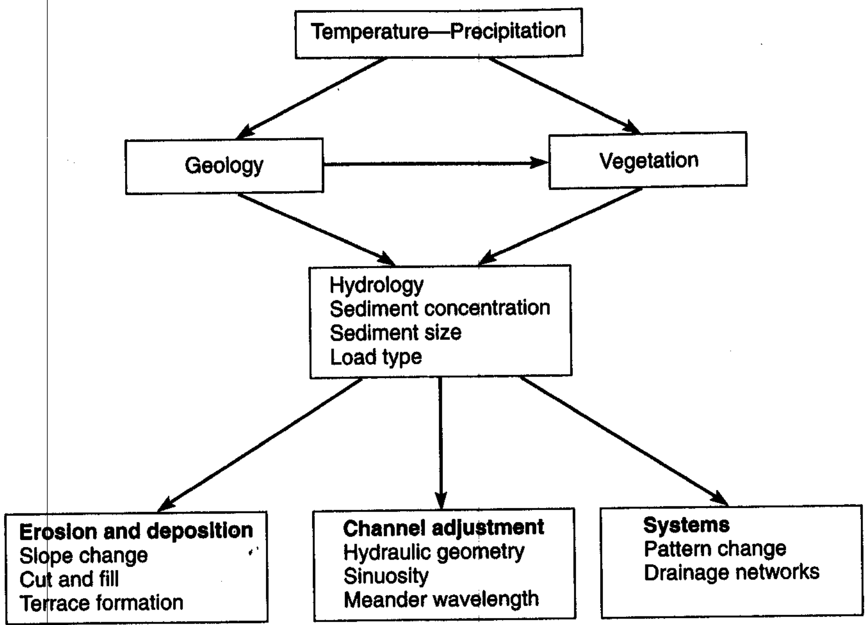


**Figure 2.16**
Hypothetical flow chart showing how climatic variables exert an influence on rivers. A change in climate alters sediment concentration, sediment size, or load type, requiring a response by the river system. Responses vary depending on local conditions.

much water and sediment get to a river channel. Variations in these two factors require threshold adjustments that can occur in a number of different ways (fig. 2.16). In addition, the lag in response is controlled by how rapidly a new vegetal screen and its characteristics can develop under a new climatic regime. This factor is not as straightforward as we would like. Predictions of response to possible future climate changes are clouded because vegetation biomes before the Holocene (ca. 10,000 years B.C.) have no modern analogs (Overpeck et al. 1992). This fact suggests that scientists cannot predict with certainty what types of vegetation communities will develop if our climate reverts back to conditions that were prevalent during the Pleistocene.

Another problem is that the same climatic change may prompt entirely different sediment yields and therefore different geomorphic responses. For example, considering figure 2.15 again, assume that a 15-cm decrease in precipitation occurs in a particular drainage basin that had an effective precipitation of 45 cm prior to the change. The reduced precipitation will result in a greatly increased sediment yield. However, the same 15-cm decrease in a basin having a 35-cm annual precipitation prior to the change will produce a major reduction in sediment yield. Theoretically, then, the same climate change may result in cutting by one river and filling by another because the type and amount of sediment yielded during adjustment to the new climate is oppositely affected. What this tells us is that the effect of climate change may be highly dependent on the antecedent values of temperature and precipitation. If that is a true statement, knowledge of preexisting climate may be as important in understanding how systems respond to climate changes as knowing the magnitude of the change itself.

In sum, the relationships among climate, process, and landform are not easily determined because the effect of change is sidetracked into ancillary factors. The adjustments of these factors in the new climate provide variable conditions of load and water that spur responses that are not predictable under the present state of our knowledge. Thus, we have not been able to describe clearcut relationships between climate and landforms because we are far from understanding the climatic geomorphology scheme.

## SUMMARY

In this chapter we briefly examined climate and endogenic factors as major external controls on geomorphic systems. Endogenic influence occurs primarily through the addition of mass and energy by volcanism and tectonic activity. The most important tectonic processes are those producing vertical movements of the surface. One of these is the isostatic adjustment required when the internal mass balance is upset. Other vertical movements, associated with faulting and warping, are integral parts of a subdiscipline known as tectonic geomorphology, in which the relationships among tectonics, processes, and landforms are utilized in a variety of geologic and environmental studies. In most cases, vertical displacements induce threshold conditions and responses in the affected surficial systems.

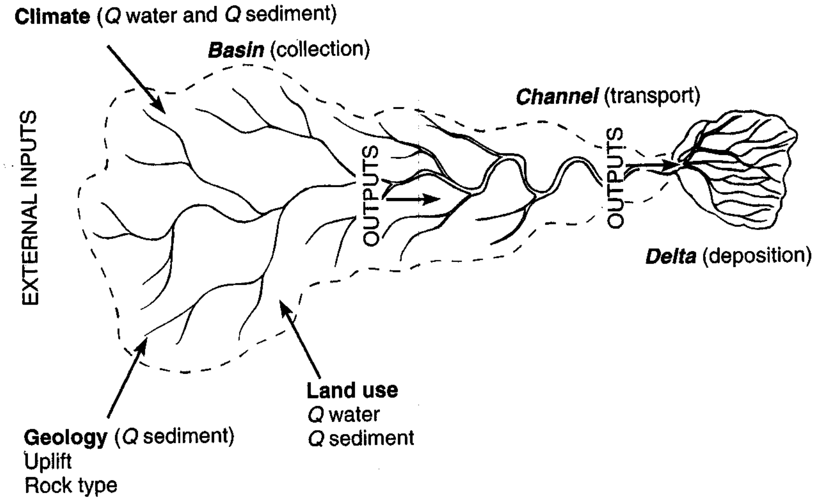


**Figure 5.1B**
Schematic surface components of the fluvial system. The tributaries provide links between lithology and climate and are adjusted to both. Channel characteristics vary in response to the external variables of sediment and water discharge ($Q$), which are influenced naturally from climate, tectonic, and lithologic factors. Human influence also modifies these variables through land use alterations.

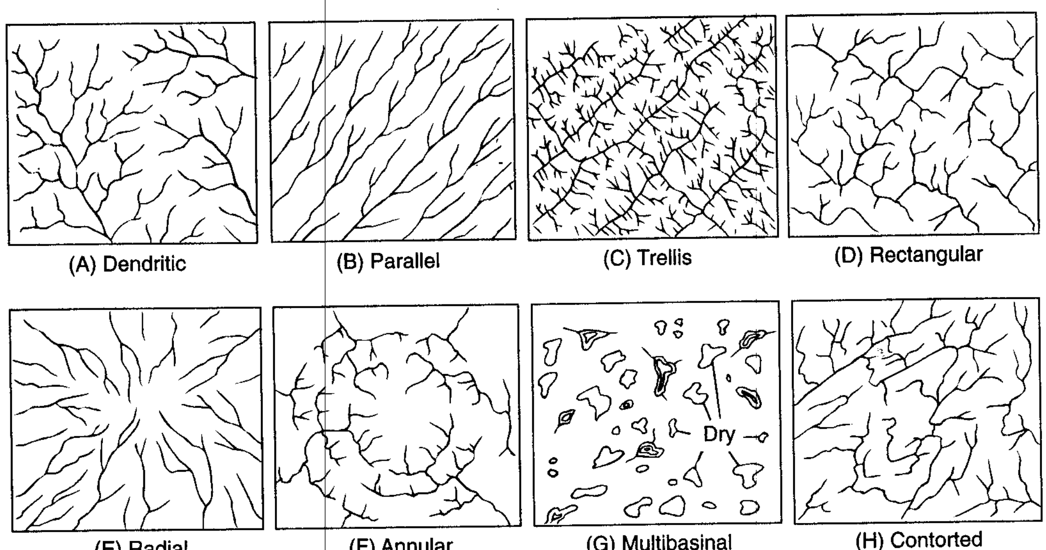


**Figure 5.2A**
Basic drainage patterns. Descriptions are given in table 5.1.
(Howard 1967, reprinted by permission)

underlying geology, described in figure 5.2A and table 5.1. Because the gross character of these patterns is evident on topographic maps and aerial photos, the patterns are useful for structural interpretation (fig. 5.2B) (Howard 1967) and for approximating lithology in a study of regional geology.

In a hydrologic sense, however, prior to World War II most basins were described in qualitative terms such as well-drained or poorly drained, or they were connoted descriptively in the Davisian scheme as youthful, mature, or old. The mechanics of how river channels or networks actually form and how water gets into a channel was poorly understood by geologists and hydrologists alike. This early twentieth-century view of streams and drainages contrasts markedly with the avant-garde approach presented by R. E. Horton during the latter part of this period (Horton 1933, 1945). His attempt to explain stream origins in mathematical terms and to describe basin hydrology as a function of statistical laws marked the birth of quantitative geomorphology. We now know that many of Horton's original ideas are only partially correct. Still, modern geomorphic analysis of drainage basins has its roots in Horton's original work, and his thinking has been instrumental in the development of modern geomorphology.

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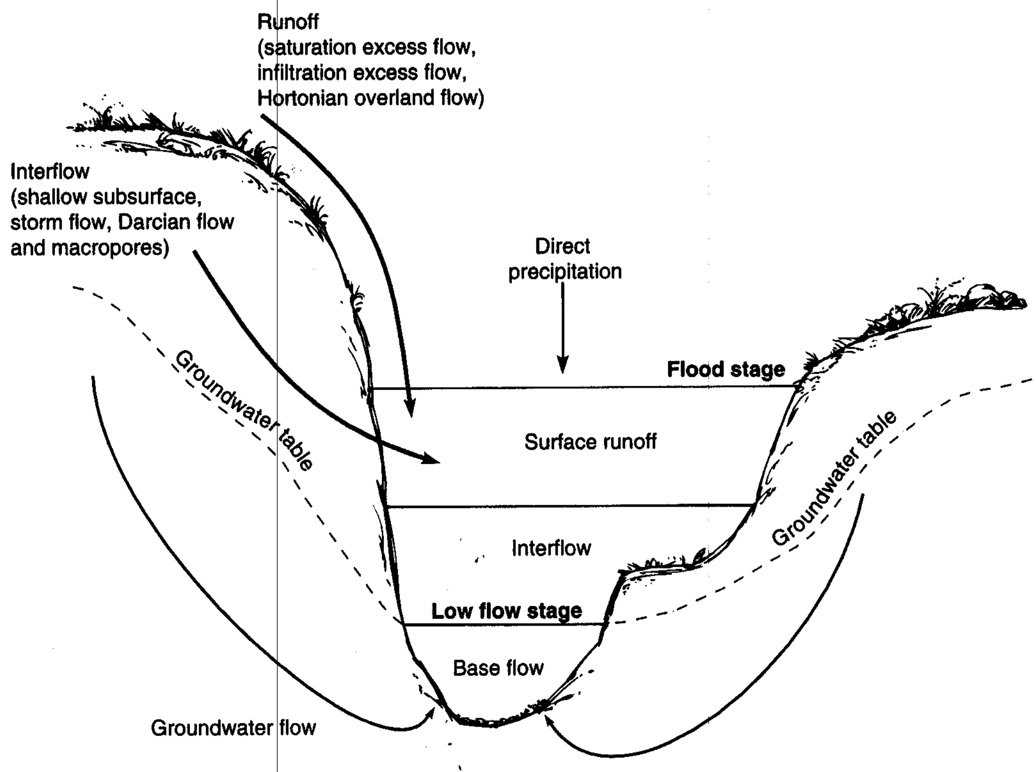


**Figure 5.7**
Schematic of stream channel showing major kinds of water contributions. Arrival of water from any given precipitation event is progressively delayed from runoff to interflow to groundwater flow.
(Based on Kochel 1992)

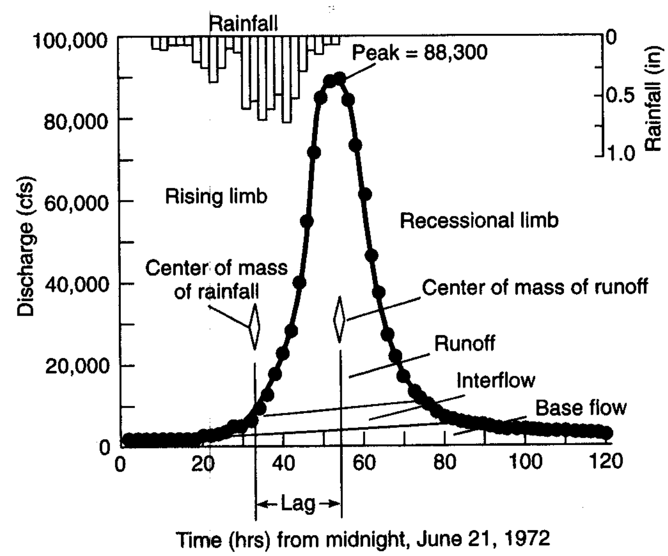


**Figure 5.8**
Flood hydrograph of the Hurricane Agnes flood of June 1972 on the Conestoga River at Lancaster, Pa. Although the curve is rather symmetrical, most hydrographs show significant skewness with a broader recessional limb reflecting interflow and groundwater inputs after a storm.

of the physical processing of precipitation from the divides to the site of measurement. Sophisticated models have been developed to predict how water is collected, stored, routed, and summed from all parts of a basin to achieve a final output hydrograph (for example, U.S. Army Corps Engineers 1985). If geology and topography are alike throughout an area, then rainfalls having similar properties should generate hydrographs with the same shape. On this premise, a *type* hydrograph for a basin, called a **unit hydrograph,** has been developed, in which the runoff volume is adjusted to the same unit value (i.e., one inch of rainfall spread evenly over the basin over one day). The unit hydrograph has been used as a connecting link in many studies attempting to relate basin morphometry to hydrology. By comparing the shapes of unit hydrographs from different basins, we can see the effects of differences in physical attributes of the basin.

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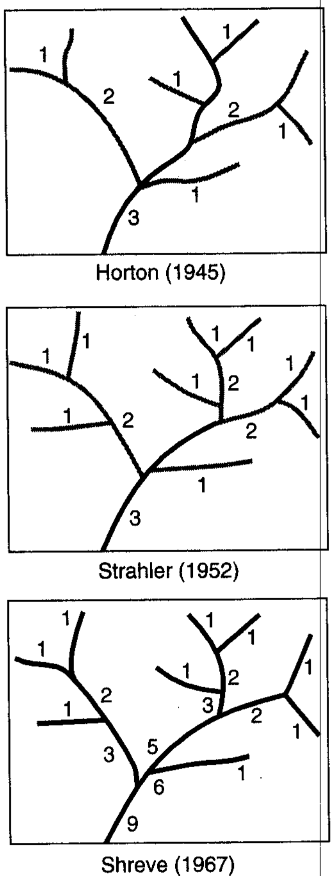


**Figure 5.17**
Methods of ordering streams within a drainage basin.

Shreve's system appears in many of the sophisticated runoff modeling packages which are beyond the scope of this discussion (for example, see Smart and Wallis 1971; Abrahams 1980; Abrahams and Miller 1982).

Every basin possesses a quantifiable set of geometric properties that define the linear, areal, and relief characteristics of the watershed (table 5.2), known as the basin morphometry. These variables correlate with stream order, and various combinations of the parameters obey statistical relationships that hold for a large number of basins. Two general types of numbers have been used to describe basin morphometry or network characteristics (Strahler 1957, 1964, 1968). *Linear scale* measurements allow size comparisons of topographic units. The parameters may include the length of streams of any order, the relief, the length of basin perimeter, and other measurements. The second type of measurement consists of *dimensionless numbers,* often derived as ratios of length parameters, that permit comparisons of basins or networks. Length ratios, bifurcation ratios, and relief ratios are common examples. Table 5.2 shows the most commonly used linear, areal, and relief equations, but numerous others have been derived from these.

**Linear Morphometric Relationships** The establishment of stream ordering led Horton to realize that certain linear parameters of the basin are proportionately related to the stream order and that these could be expressed as basic relationships of the drainage composition. Much of linear morphometry is a function of the *bifurcation ratio* ($R_b$), which is defined as the ratio of the number of streams of a given order to the number in the next higher order (using Strahler ordering). The bifurcation ratio allows rapid estimates of the number of streams of any given order and the total number of streams within the basin. Although the ratio value will not be constant between each set of adjacent orders, its variation from order to order will be small, and a mean value can be used. Also, as Horton pointed out, the number of streams in the second highest order is a good approximation of $R_b$. When geology is reasonably homogeneous throughout a basin, $R_b$ values usually range from 3.0 to 5.0.

The *length ratio* ($R_L$), similar in context to the bifurcation ratio, is the ratio of the average length of streams of a given order to those of the next higher order. The length ratio can be used to determine the average length of streams in an unmeasured given order ($L_O$) and their total length. The combined length of all streams in a given basin is simply the sum of the lengths in each order. For most basin networks, stream lengths of different orders plot as a straight line on semilogarithmic paper (fig. 5.18), as do stream numbers. The relationships between stream order and the number and length of segments in that order have been repeatedly verified and are now firmly established (Schumm 1956; Chorley 1957; Morisawa 1962; and many others).

**Areal Morphometric Relationships** The equity among linear elements within a drainage system suggests that areal components should also possess a consistent morphometry, because dimensional area is simply the product of linear factors. The fundamental unit of areal elements is the area contained within the basin of any given order ($A_O$). It encompasses all the area that provides runoff to streams of the given order, including all the areas of tributary basins of a lower order as well as interfluve regions. Schumm (1956) demonstrated (fig. 5.19) that basin areas, like stream numbers and lengths, are related to stream order in a geometric series.

Although area by itself is an important independent variable (Murphey et al. 1977), it has also been employed to manifest a variety of other parameters (see table 5.2), each of which has a particular significance in basin geomorphology, especially in regard to the collection of rainfall and concentration of runoff. Numerous studies have been successful in formulating relationships between basin area and discharge. One of the more important areal factors is *drainage density* ($D$), which is

# FLUVIAL FIELD TRIP EQUATION LIST

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## FROUDE NO. – DESCRIBES FLOW TYPE

$$Fr = \frac{V}{\sqrt{dg}}$$

V = velocity m/sec
d = depth m
g = gravity acc. = 9.8 m/sec²

- F < 1 = TRANQUIL FLOW
- F = 1 = CRITICAL FLOW
- F > 1 = SUPER CRITICAL FLOW

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## MANNING'S EQUATION – TO CALCULATE STREAM VELOCITY

$$V = \frac{R^{2/3} S^{1/2}}{n}$$

V = VELOCITY m/sec
R = HYDRAULIC RADIUS = $\frac{A}{P}$ — A = CHANNEL AREA m², P = WETTED PERIMETER m
S = SLOPE
n = ROUGHNESS

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$$\tau_c = \gamma R S$$ = CRITICAL SHEAR FORCE FOR EROSION

γ = SPECIFIC WT. OF $H_2O$ = 9800 N/m³
R = HYDRAULIC RADIUS = $\frac{A}{P}$
S = SLOPE

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## STREAM POWER = KINETIC ENERGY AVAILABLE FOR WORK

TOTAL POWER $\Omega = \gamma Q S$ (WATTS)

UNIT POWER $\omega = \frac{\gamma Q S}{w}$ (WATTS/m)

γ = SP. WT. $H_2O$ = 9800 N/m³
Q = DISCHARGE = m³/sec
S = SLOPE
w = width m

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## CONTINUITY EQUATION

$$Q = VA$$

Q = DISCHARGE m³/sec
V = VELOCITY m/sec
A = AREA OF CHANNEL m²

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## Water Budget

$$P = R + I + S$$

P = Total Precipitation

R = Surface Runoff

I = ~~Soil~~ Interception

S = Soil/Groundwater Storage

## Recurrence Interval

RI = Total No. of Years Record / No. Observations in Time Frame

## Fluid Mechanics

Force of Water

$$F_p = F_g(\sin\theta)$$

$f_p$ = force parallel to channel

$f_g$ = force of gravity

$\theta$ = angle of channel slope

Momentum

$m$ = (mass)(velocity)

Froude No.

$$Fr = \frac{u}{\sqrt{dg}}$$

u = velocity m/s, g = force of gravity m/s²

d = depth of flow m

Reynolds No.

$$Re = \frac{\rho d u}{v}$$

ρ = fluid density, v = viscosity, d = depth, u = velocity

Re = Reynolds No.

## Porosity & Permeability

$$P\% = \frac{V_T - V_S}{V_T} \times 100\%$$

$V_T$ = total vol.

$V_S$ = vol. solid

## Darcy's Law

$$Q = \frac{KA(P_2 - P_1)}{\mu L}$$

Q = Discharge ($L^3/T$)

K = Permeability

A = x-sectional area

L = length

$P_2 - P_1$ = Head

μ = viscosity

## River Morphometry

- Sinuosity

$$S = \frac{\text{Stream length}}{\text{valley length}}$$

- Stream Stream orders
- Drainage Frequency $f = \frac{No.}{A}$
- Bifurcation Ratio

$$R_b = \frac{N_o}{N_{o+1}}$$

- Length Ratio

$$R_L = \frac{L_{AVG\,o}}{L_{AVG\,o+1}}$$

- Drainage Density

$$D = \frac{L_{TOTAL}}{A}$$

- Texture Ratio

$$T = N/P$$

## Stream Flow

Continuity Equation

$$Q = VA = Vwd$$

Q = Discharge ($L^3/T$)

V = Velocity ($L/T$)

A = x-sectional area

w = channel width

d = channel depth

$$P = 2d + w$$

P = wetted perimeter

Hydraulic Radius

$$R = A/P = m^2/m$$

Manning Equation

Metric Units

$$V = \frac{R^{0.66} S^{0.5}}{n}$$

V = mean velocity m/s

n = coefficient of roughness

R = hydraulic radius m

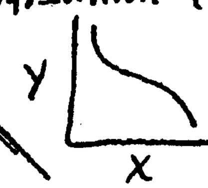
S = slope (gradient)

## Hypsometric Analysis

$$x = \frac{a}{A} = \frac{\text{area above datum}}{\text{total basin area}}$$

$$y = \frac{h}{H} = \frac{\text{Ht. above datum}}{\text{total basin relief}}$$

Hypsometric Curve

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[redacted]

# TABLE 1 – ROUGHNESS

A. [redacted] Values of Roughness, n

| River Description | Roughness, n |
|---|---|
| **Ordinary rivers:** | |
| clean, straight channel, no riffles or pools | 0.030 |
| straight, weedy, boulders | 0.035 |
| clean winding channel, pools and riffles | 0.040 |
| weedy, winding, deep pools | 0.070 |
| **Alluvial channels:** | |
| vegetated, no brush, grassy | 0.030–.035 |
| vegetated, brushy | 0.050–.10 |
| no vegetation | |
| ripples, dunes | 0.017–.035 |
| plane bed | 0.011–.015 |
| antidunes | 0.012–.020 |
| **Mountain streams: rocky beds** | |
| no vegetation, steep banks | |
| bed of gravel, cobbles, | 0.040 |
| bed of cobbles and boulders | 0.050 |

Compiled and adapted from Chow (1959 and 1964)

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B.

[redacted] 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 |

Source: *Handbook of Applied Hydrology*, ed. by Ven T. Chow, copyright 1964 McGraw-Hill Publishing Co., Inc.

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# HYDROLOGY SURFACE WATER EQUATION LIST

(1) CONTINUITY EQUATION FOR RIVER DISCHARGE

$$Q = AV = wdV = wd\frac{L}{t} = \frac{Vol}{t}$$

Where $Q$ = DISCHARGE ($L^3/t$)
$A$ = Channel cross-sectional AREA ($L^2$)
$V$ = VELOCITY ($L/t$)
$w$ = CHANNEL WIDTH ($L$)
$d$ = CHANNEL DEPTH ($d$)
$Vol$ = VOLUME ($L^3$)
$t$ = time (time units)

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(2) WATERSHED DRAINAGE DENSITY

$$D_d = \frac{\Sigma L}{A_d}$$

where $\Sigma L$ = SUM OF TOTAL STREAM LENGTHS ($L$)
$A_d$ = drainage Area ($L^2$)
$D_d$ = DRAINAGE DENSITY ($L/L^2 = m/km^2$)

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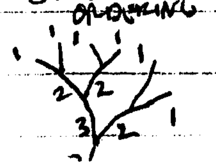
(3) SHREVE MAGNITUDE FOR WATERSHEDS

STREAM ORDERING




$M = \Sigma$ FREQUENCY OF FIRST ORDER STREAMS

(4) RATIONAL RUNOFF METHOD (for WATERSHEDS)

$$Q_p = CIA \quad \text{where}$$

$Q_p$ = PEAK RUNOFF DISCHARGE ($L^3/t$)
C = RATIONAL RUNOFF COEFFICIENT (dimensionless)
I = RAINFALL INTENSITY ($L/t$)
$A_d$ = DRAINAGE AREA ($L^2$)

VALUES OF "C"

PAVEMENT C = 0.70 – 0.95
SANDY SOILS C = 0.20 – 0.40
CLAYEY SOILS/COLLUVIUM C = 0.40 – 0.50

NOTE: WHERE SOILS ARE 100% SATURATED, $C \longrightarrow 1.0$,
SO IN THIS CASE $Q_p = IA$

(5) FLOOD RECURRENCE INTERVAL

$$R.I. = \frac{n+1}{m}$$

where n = TOTAL NO. OF EVENTS OR YEARS
m = RANK OF EVENT, WITH #1 = LARGEST

$$P = \frac{1}{R.I.}$$

P = PROBABILITY OF GIVEN MAGNITUDE OF FLOOD

(6) PEAK DISCHARGE

$Q_p$ = MAXIMUM DISCHARGE ON RECORD ($L^3/t$)

$Q_p$ DAILY = MAX. DAILY DISCHARGE

$Q_p$ ANNUAL = MAX YEARLY Q

(7) EMPIRICAL HYDROLOGIC RELATIONS FOR SELECT REGIONAL WATERSHEDS

(A) $Q_{max} = 38 M^{0.89} D^{-0.50}$ FOR APPALACHIAN PLATEAU REGION

where $Q_{max}$ = maximum DISCHARGE ($L^3/t$)
M = STREAM MAGNITUDE (DIMENSIONLESS)
D = DRAINAGE DENSITY ($L/L^2$)

(B) $Q_{2.33} = 34.5 A^{0.93}$ (VERMONT WATERSHED)

$Q_{2.33}$ = DISCHARGE WITH A 2.33 yr RECURRENCE INTERVAL

A – DRAINAGE AREA

(C) GENERALIZED RELATIONSHIP

$Q_x = a A^b$ b range: 0.5–0.9

where x = RECURRENCE INTERVAL, Q = DISCHARGE, A = DRAINAGE AREA, a = Coefficient, b = exponent.

(8) TIME FOR HYDRAULIC CONCENTRATION OF DRAINAGE BASIN

↓

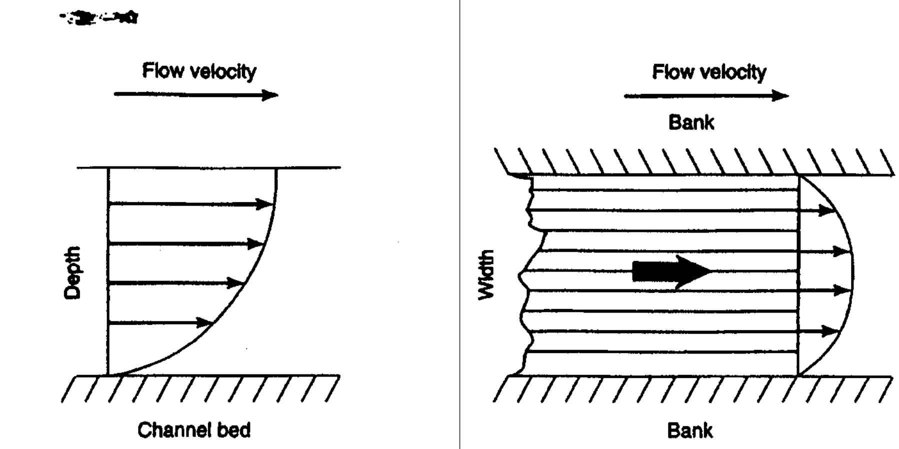
DEFINED: TIME REQUIRED DURING A STORM, FOR OVERLAND AND CHANNEL FLOWS TO TRAVEL FROM THE MOST DISTANT DRAINAGE DIVIDE TO THE OUTLET OF THE BASIN

$$t_c = \frac{L^{1.15}}{7700\, H^{0.38}}$$ (EMPIRICAL EQUATION)

$t_c$ = time of concentration (hours)

$L$ = LENGTH FROM DIVIDE TO BASIN OUTLET (ft)

$H$ = BASIN RELIEF BETWEEN DIVIDE AND OUTLET (ft)




(A) (B)

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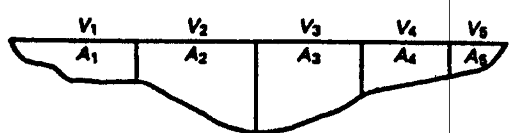
**FIGURE 6.1**
Diagram showing the changes in flow velocity with (A) flow depth and (B) flow width. Resistance to flow along the bed and banks allows the greatest velocities to occur toward the center of the channel near the water surface.




subareas of velocity domains.

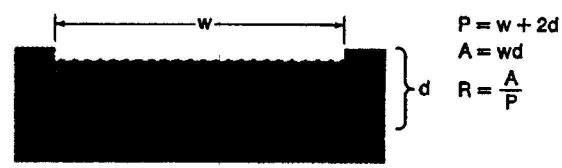



**FIGURE 6.2**
Cross-sectional measurements of a stream channel: *w* = width, *d* = depth, *A* = area, *R* = hydraulic radius, *P* = distance along wetted perimeter.

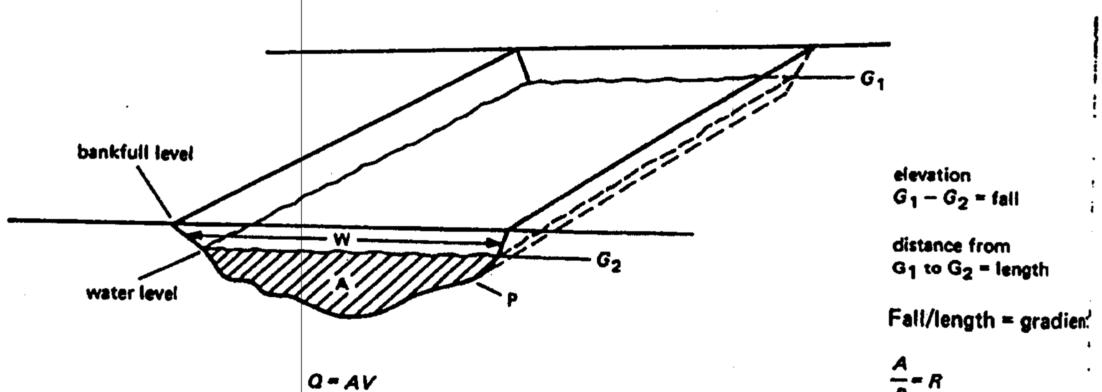



*Figure 9.2.* Nomenclature of channel morphology.

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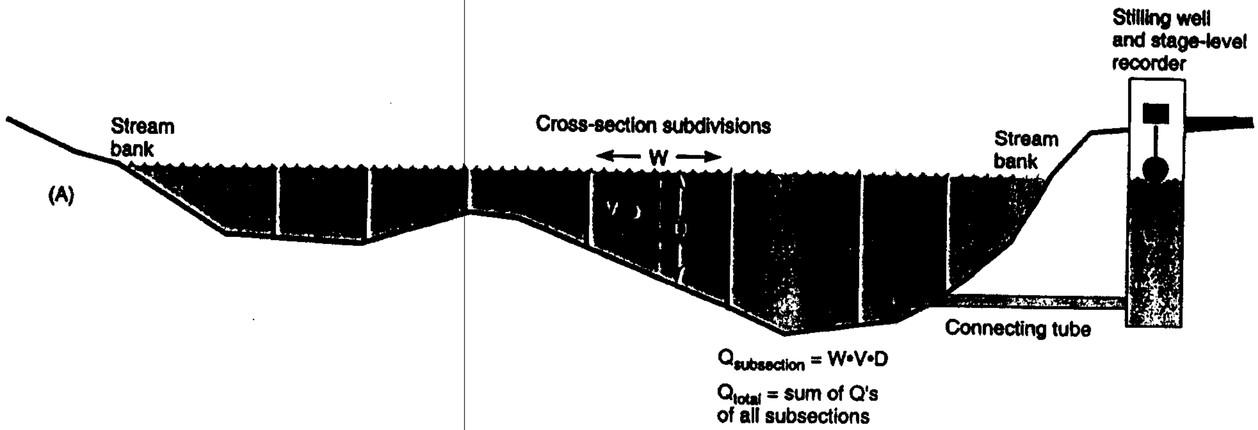


(D)

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**FIGURE 5.33**
Rating curve for low flow. Rock Creek near Red Lodge, Mont.

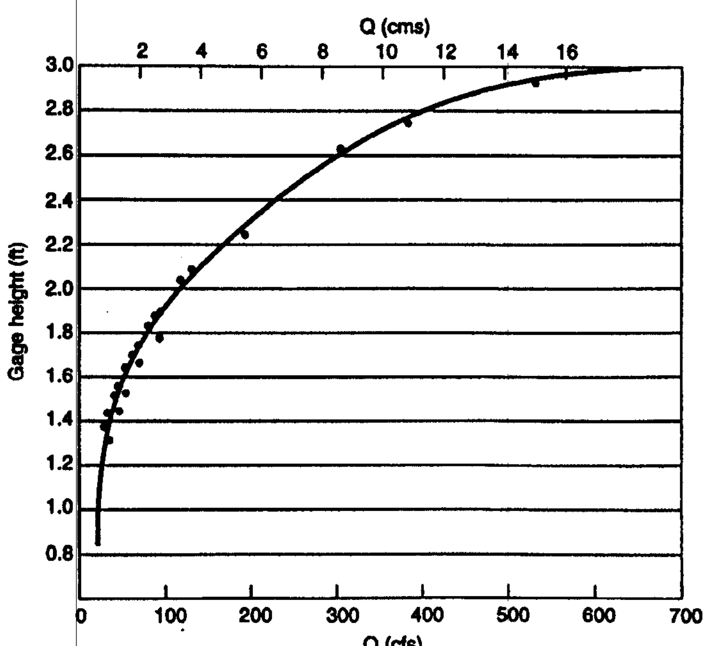


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**TABLE 6.1 Flow Characterization in Open Channels.**

| 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 |

another (Leopold et al. 1964). The intensity of the resistance is related to the *molecular viscosity* of the fluid, where viscosity is governed by internal characteristics of the fluid such as temperature and the concentration of suspended sediment.

In **turbulent flow,** the water does not move in parallel layers; its velocity fluctuates continuously in all directions within the fluid. Water repeatedly interchanges between neighboring zones of flow, and shear stress is transmitted across layer boundaries in another form of viscosity, called *eddy viscosity.* Eddy viscosity greatly increases the flow resistance and thus the dissipation of energy. Because turbulence is generated along the channel boundaries, most resistance in this type of flow results from external factors such as the channel configuration and the size of the bed material.

As depth and velocity increase, the conditions at which laminar flow changes to turbulent can be predicted by a dimensionless parameter called the **Reynolds number (Re):**

$$Re = VR\rho/\mu$$

where $V$ is the mean velocity, $R$ the hydraulic radius, $\rho$ the density, and $\mu$ the molecular viscosity. The hydraulic radius is determined by the relationship

$$R = A/P$$

where $A$ is the cross-sectional area of the channel and $P$ is the wetted perimeter (fig. 6.1). In wide, shallow channels the hydraulic radius closely approximates the mean depth.

Because the factor $\mu/\rho$ defines the fluid property called *kinematic viscosity* ($\upsilon$), the Reynolds number represents a ratio between driving and resisting forces:

$$\mathrm{Re} = VR\rho/\mu = VR/\upsilon = \frac{\text{driving forces}}{\text{resisting forces}}$$

In normal situations true laminar flow occurs where $Re$ values are less than 500, and well-defined turbulent flow when $Re$ is greater than about 2000.

Another dimensionless number used to describe the conditions of flow is the **Froude number (Fr):**

$$Fr = V/\sqrt{dg}$$

where $d$ is depth and $g$ is gravity. The Froude number is important because it can be used to distinguish subtypes of turbulent flow called *tranquil flow* ($Fr < 1$), *critical flow* ($Fr = 1$), and *rapid flow* ($Fr > 1$). The energy that is expended by these flow types differs considerably. In addition, within sand bed channels, tranquil, critical, and rapid flow have been related to the development of distinct sedimentary bedforms (fig. 6.2), which also exert an important influence on the resistance to flow in open channels (as will be discussed in more detail in the next section).

Flow within natural channels is invariably turbulent, although a very thin layer of quasi-laminar flow may be present along the channel boundaries. Most of the turbulence is generated along the water and sediment interface, causing an increase in resistance and a decrease in velocity toward the channel perimeter (fig. 6.3). Thus, across a channel the highest velocities occur near the center of the flow. The location of highest velocities may vary significantly, however, as a function of channel alignment and cross-sectional shape (fig. 6.3B), becoming more asymmetrical in meander bends (Knighton 1998).

In rivers formed in sand or finer-grained sediments with smooth channel beds, the vertical velocity profile is typically characterized by two zones of flow in addition to the laminar sublayer (fig. 6.3A). The lower zone encompasses about 20 percent of the total flow depth, and exhibits a quasi-logarithmic decrease in velocity toward the channel floor. The overlying upper zone is less affected by flow resistance along the channel bed, and vertical velocity profiles are more nearly parabolic in

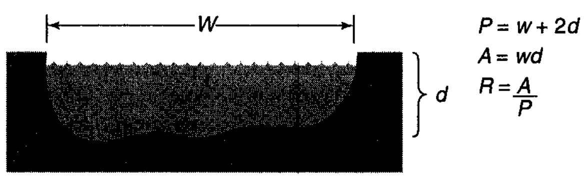


**Figure 6.1**
Cross-sectional measurements of a stream channel: $w$ = width, $d$ = depth, $A$ = area, $R$ = hydraulic radius, $P$ = distance along wetted perimeter.

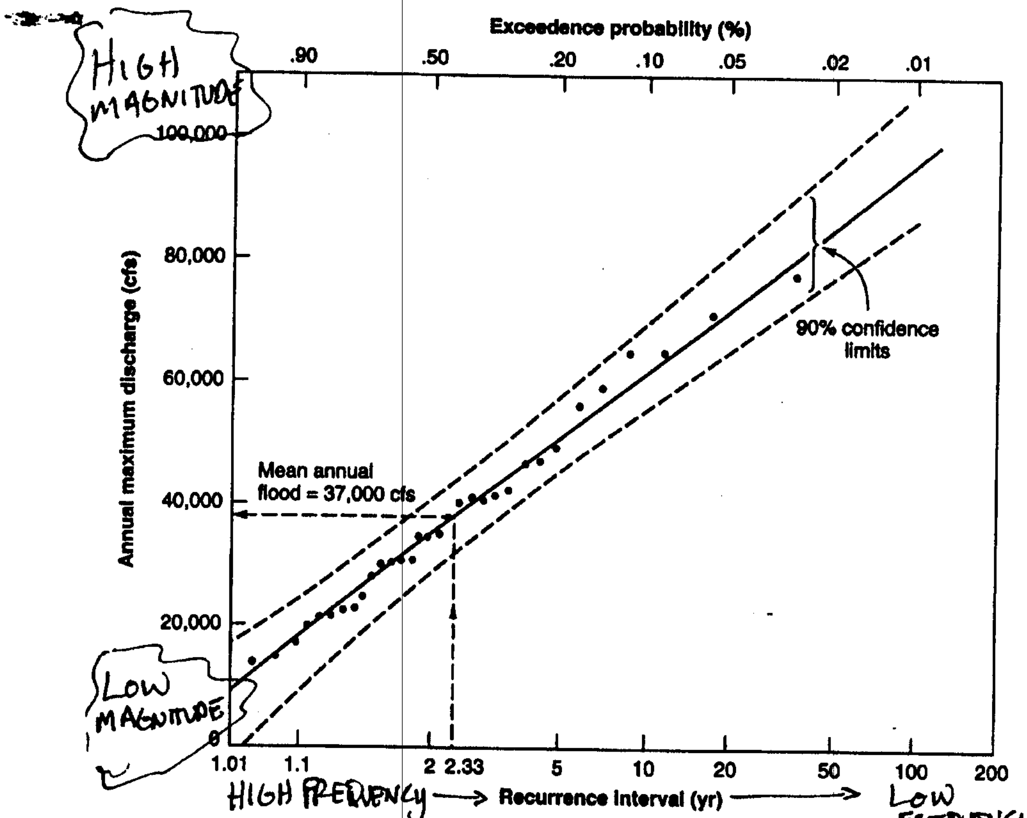


**FIGURE 5-8**
Flood frequency curve for annual floods on the Skykomish River, Washington. Dashed lines are 90-percent confidence limits. (Data from U.S. Geological Survey; plot from Dunne and Leopold, 1978)

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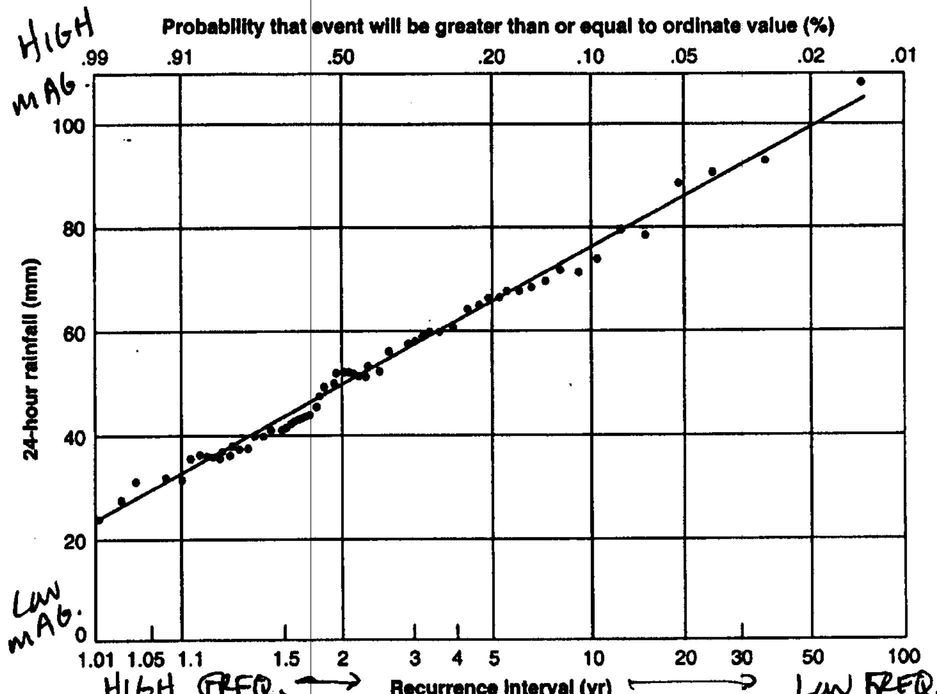


**FIGURE 5-1**
Recurrence interval of 24-hour precipitation, Buffalo, New York, 1891–1961. (From Dunne and Leopold, 1978)

74H

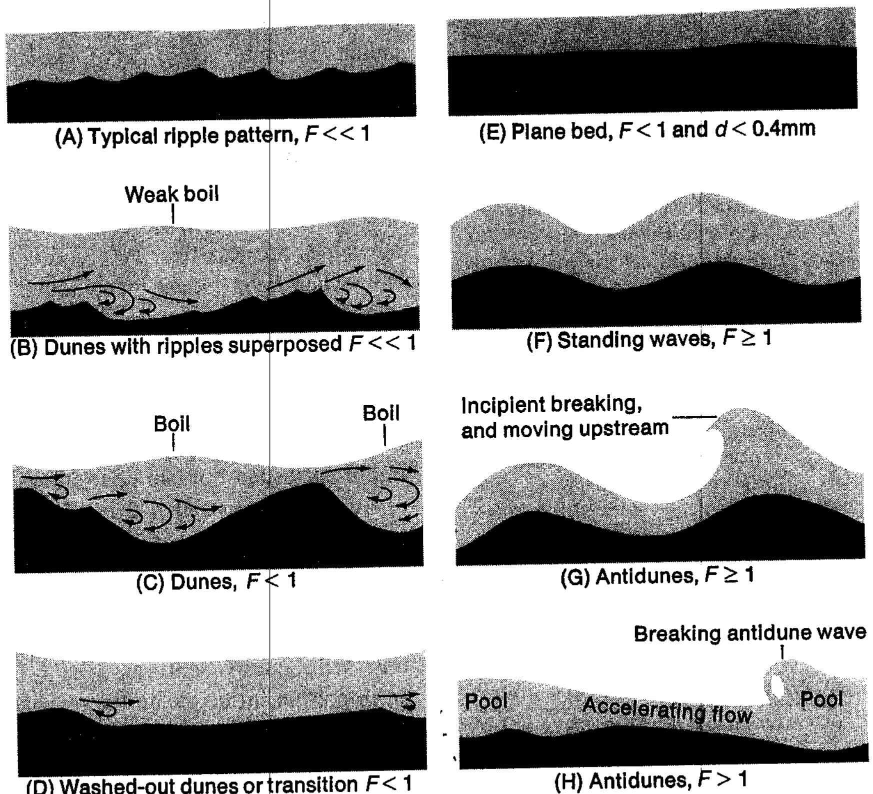


**Figure 6.2**
Bed forms in alluvial channels and their relation to flow conditions. $F$ = Froude number, $d$ = depth.
(Simmons and Richardson 1963)

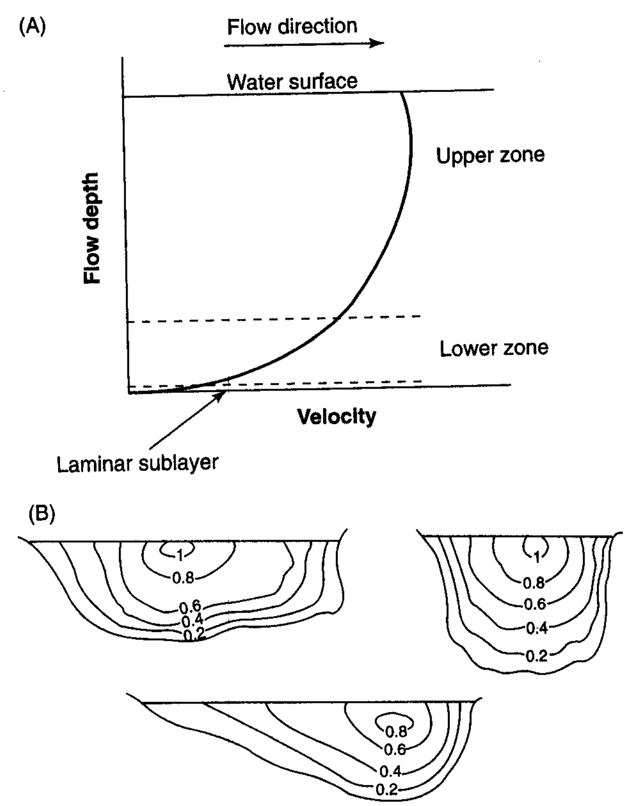


**Figure 6.3**
Variations in flow velocity as a function of water depth. The lower zone exhibits a quasi-logarithmic form induced by resistance along the channel bed. The upper zone is less affected by the bed roughness, and is more nearly parabolic in shape. The laminar sublayer may be absent or discontinuous in coarse-grained channels. (B) Typical variations in velocity across the channel. Isovels (lines of equal velocity) are in m/s.
(Modified from Wolman 1955)

shape (Wiberg and Smith 1991). For the purposes of calculating discharge, it is often assumed that the velocity profile exhibits a logarithmic form through the entire depth of flow (Wiberg and Smith 1991; Pitlick 1992) and that the average velocities are found at 0.6 of the depth from the water surface (Byrd et al. 2000). The highest velocities exist at, or immediately below, the surface (fig. 6.3). In rough, coarse-grained rivers where the fluid is forced around large clasts, the vertical velocity profile may become irregular or distorted (Bathurst 1988; Wiberg and Smith 1991). In fact, the lower zone of flow may be completely absent (Wiberg and Smith 1991). These effects appear most pronounced where flow depths are small compared to the size of the roughness elements (including large clasts and other topographic features) on the channel floor. Byrd et al. (2000) suggest that for these rivers, the average velocities used in the calculation of discharge may be obtained most effectively by averaging two or three measurements obtained at differing depths.

## Flow Equations and Resisting Factors

Flow and resistance have been the concern of hydraulic engineers for centuries, and a number of equations have been derived to express the relationships between the two factors. Two equations of great importance to students of rivers are the **Chezy equation** and the **Manning equation.** Both were derived from equating driving and resisting forces in nonaccelerating flow, and both have been employed in a variety of fluvial investigations. Derived in 1769, the Chezy equation

$$V = C\sqrt{RS}$$

shows that velocity is directly proportional to the square root of the $RS$ product, where $S$ is slope of the channel. The Chezy coefficient ($C$) is a constant of proportionality that is related to resisting factors in the system.

The Manning equation originated in 1889 from an attempt by Manning to systematize the existing data into a useful form. The equation, when utilizing English units, is expressed as

$$V = \frac{1.49}{n} R^{2/3} S^{1/2}$$

and is similar to the Chezy formula in that velocity is proportional to $R$ and $S$. In addition, the factor $n$, called the Manning roughness coefficient, is also a resisting element that is closely related to the Chezy coefficient because as

$$C(RS)^{1/2} = \left(\frac{1.49}{n}\right) R^{2/3} S^{1/2}$$

then

$$C = \frac{1.49 R^{1/6}}{n}$$

**TABLE 6.2 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 |

Source: *Handbook of Applied Hydrology*, ed. by Ven T. Chow, copyright 1964 McGraw-Hill Publishing Co., Inc.

Manning's $n$ is presumed to be a constant for any particular channel framework; consequently, it has been used extensively in analyses of river mechanics (table 6.2). The U.S. Geological Survey, for example, has developed a visual guide for rapid estimation of Manning's $n$ along any given stream reach (Barnes 1968). In reality, however, resistance coefficients vary with flow stage, channels typically becoming more hydraulically efficient as discharge increases (Bathurst 1982; Knighton 1998).

It is important to recognize that resistance is not directly measured from the flow. Rather, the resistance coefficients are defined by hydraulic characteristics (e.g., $S$, $R$, $V$) and must be indirectly estimated from measurements of the defining parameters. Moreover, they are dependent on other factors (not included in the defining equations) because in alluvial channels their values vary with particle size, sediment concentration, and bottom configuration. The coefficients, then, represent the total resistance to flow that originates from a variety of sources. Attempts to separate the total resistance into specific source types have met with only limited success. Bathhurst (1993) suggests, however, that total resistance can be subdivided into three major components: free surface, channel, and boundary resistance. *Free surface resistance* represents the loss of energy resulting from the disruption of the water by surface waves and abrupt changes in water surface gradients (hydraulic jumps). *Channel resistance* is that which is associated with undulations in the channel bed and banks as well as alterations in channel plan form and cross-sectional shape. Most studies to date have focused

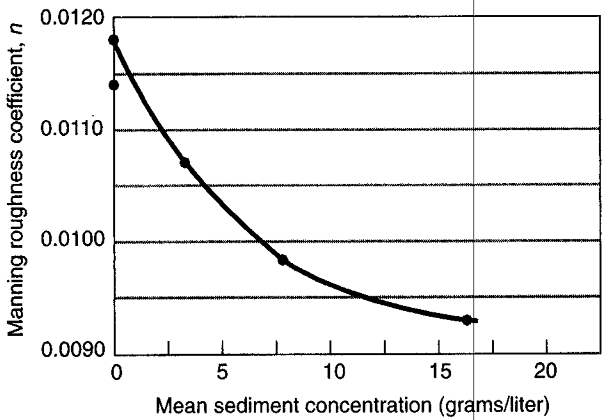


**Figure 6.5**
Effect of suspended load concentration on the Manning roughness coefficient *n*.
(Vanoni 1946)

energy causing flow. In contrast, sediment concentration (the amount of sediment per unit volume of water) internally affects resistance. This modification was first detailed by Vanoni (1941, 1946), who showed that an increase in the concentration of suspended sediment tends to lower resistance (fig. 6.5). As the concentration increases, the turbulent effect presumably is reduced because the mixing process within the fluid is dampened. All other factors being equal, sediment-laden water should flow at a higher velocity than clear water.

## SEDIMENT IN CHANNELS

Most energy in a stream is dissipated by the many factors resisting flow in open channels. The remainder, although commonly small, is used in the important task of eroding and transporting sediment. These processes, often taken for granted, are extremely complex and poorly understood, yet they underlie some of our most basic concepts of river mechanics. We will briefly review the more significant ideas about the movement of sediment in rivers.

### Transportation

In general, fine-grained sediment (silt and clay) is transported within the water column by the supporting action of turbulence. *Suspended load* usually moves at a velocity slightly lower than that of the water and may travel directly from the place of erosion to points far downstream without intermittent stages of deposition. Coarse particles may also travel in true suspension, but they are likely to be deposited more quickly and stored temporarily or semipermanently within the channel. Except for short spasms of suspension, coarse sediment usually travels as bedload. *Bedload* refers to sediment transported close to or at the channel bottom by rolling, sliding, or bouncing. How long coarse debris remains stationary within a channel depends on a large number of parameters including the nature of the debris (e.g., its size, shape, and density), the interlocking relationships between the particles, the exposure to flow, and the flow characteristics of the river; such debris probably is immobile more than it is in motion. Bradley (1970), for example, showed that gravel can be stored in channel bars long enough for weathering to drastically weaken its resistance to abrasion.

Because of fluctuating discharge, at any given time a single particle may be part of either the bedload or the suspended load. As this makes the distinction between the two load types unclear, other terms have been devised to relate sediment more appropriately to river flow. *Wash load* consists of particles so small that they are essentially absent on the streambed. In contrast, *bed material load* is composed of particle sizes that are found in abundant amounts on the streambed (Colby 1963). While most, if not all, bedload is bed material load, most bed material load is transported as suspended load.

The relationship between wash load and discharge is poorly defined because most streams at any given flow can carry more fine-grained sediment than they actually do. The concentration of fines is a function of supply rather than transporting power; therefore, it is relatively independent of flow characteristics. Coarse sediment, on the other hand, is usually available in amounts greater than a stream can carry, and so its concentration should correlate more significantly with the parameters of flow such as depth and velocity. The problem, however, is that direct measurement of bedload is extremely difficult because handheld instruments can sample for only short periods, and when they are placed on the channel bottom the flow regime is disrupted. In addition, where bedload has been continuously measured, the amount of sediment passing a given channel cross-section varies significantly with time (see, for example, Leopold and Emmett 1977; Hoey 1992; Carling et al. 1998). Furthermore, the amount of bedload at any given time varies drastically in different subwidths of the channel cross-section.

Because of the difficulties surrounding direct measurement, most estimates of bedload discharge are made by means of empirical equations that attempt to determine the maximum amount of sediment that a stream can carry (its capacity) for a given set of channel, sediment, and flow conditions (Meyer-Peter and Muller 1948; Einstein 1950; Bagnold 1980; Parker et al. 1982; Williams and Julien 1989). These equations, however, are themselves problematical; their accuracy is difficult to assess because reliable measurements of bedload discharge are scarce, and variations in bedload transport for any given set of hydraulic conditions can be large. In fact, Gomez and Church (1989) compared 10 transport formulas and concluded that for coarse-grained streams,

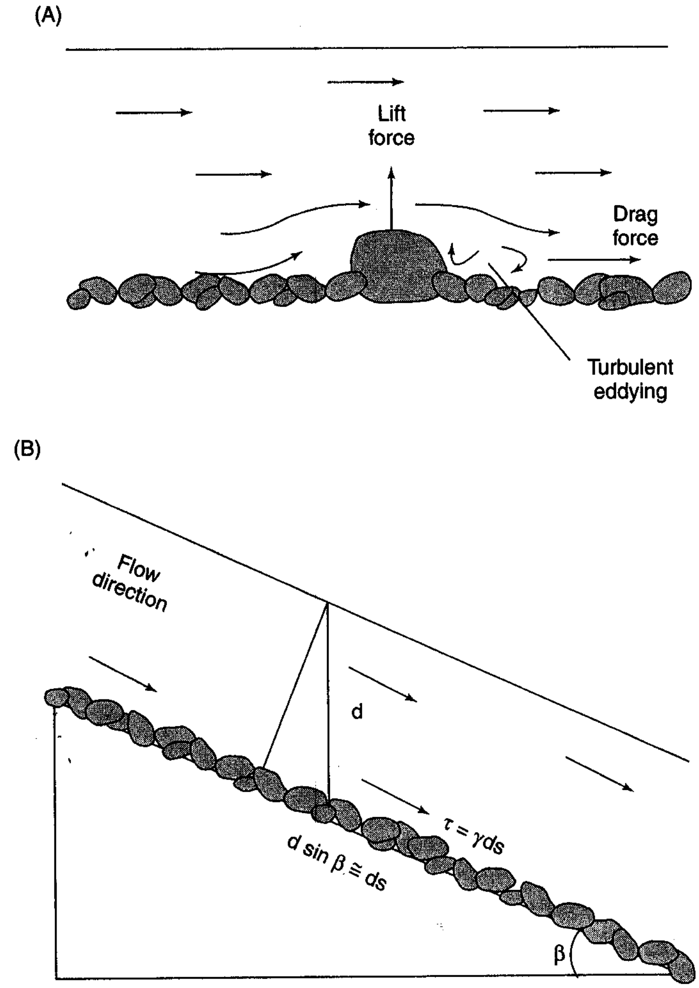


**Figure 6.8**
(A) Orientation of lift and drag forces acting on submerged channel bed sediment. Lift forces are due to variations in flow velocity over the top and bottom of the particle. Turbulent eddying may also create upward directed forces that act on the particles. (B) Component of flow weight exerted as shear stress on the channel bottom. The critical shear stress is equal to the depth-slope product (dS) multiplied by the specific weight of the water $\gamma$ and $\beta$ is the angle of slope.

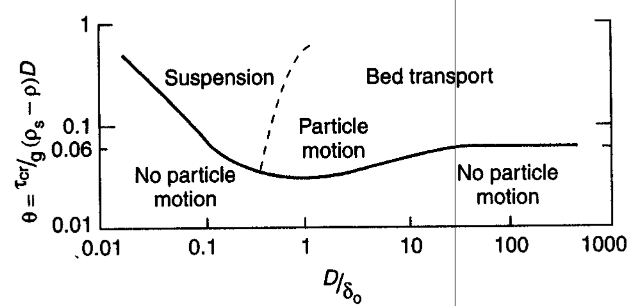


**Figure 6.9**
Shield curve for the entrainment of bed particles where $D$ is grain diameter, $\tau_{cr}$ is critical shear stress, $\rho_s$ is sediment density, $\rho$ is fluid density, and $\delta_o$ is thickness of laminar sublayer.

The Shields diagram illustrates that within hydraulically smooth channels characterized by silt and clay, dimensionless shear stress ($\theta$) varies with grain Reynolds numbers ($D/\delta_o$), reaching a minimum at a value of $D/\delta_o$ of approximately 0.03 (fig. 6.9). Dimensionless shear stress increases for smaller values of $D/\delta_o$ (fig. 6.9). Given that grain Reynolds number is related to particle size, it follows that more shear stress is required to entrain fine-grained sediments that reside below the surface of the laminar sublayer and that are not subjected to the effects of turbulent flow. Cohesion, generally associated with smaller particles, may also play a role in increasing the shear stress required for entrainment. For hydraulically rough channel beds (in which the particles are relatively large in comparison to the thickness of the laminar sublayer), motion is initiated predominantly by turbulent action (Morisawa 1985), and $\theta$ obtains a constant value of approximately 0.06 (although constant values as low as 0.03 have been reported in some studies).

Knighton (1998) notes that a disadvantage of critical shear stress formulas is that they ignore the effects of lift that may promote particle entrainment. Lift is primarily generated by differences in the velocity of the flow over the top and bottom of an individual particle, a process that creates a vertical pressure gradient leading to the upward motion of the grain (fig. 6.8). Lift may also be created by turbulent eddying generated downstream of the particle that produces locally upward directed flow. The use of critical shear stress in competence studies has been criticized for other reasons as well (Yang 1973), but the simple reality that depth and

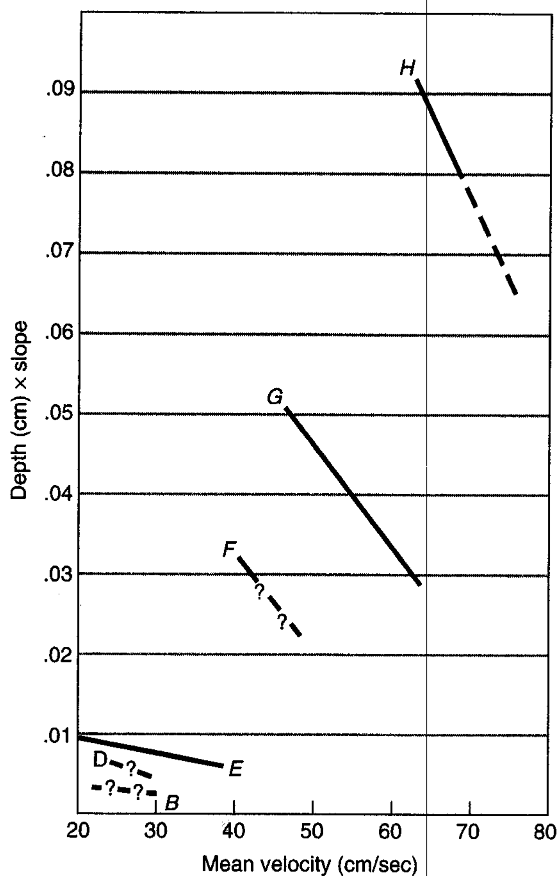


**Figure 6.10**
Sediment particles of different sizes begin to move on the streambed at different values of mean velocity and depth-slope. Smallest particles (*B*) move mainly as a function of *dS* while largest particles (*H*) move primarily as a function of the mean velocity.
(Rubey 1938)

slope in a river are easier to measure than bed velocity makes it an appealing parameter.

Precisely how the two methods correlate with each other is not completely understood, but both approaches suggest that it is more difficult to entrain particles that are either smaller or larger than medium sand. This helps explain the commonly observed phenomenon of sand-sized debris being transported across stationary material of a smaller size. There is, however, some evidence to suggest that the importance of shear stress and flow velocity to the entrainment process may vary with particle size. Rubey (1938), for example, suggested that critical bed velocity becomes more important in the entrainment process as particle size increases from fine sand to pebbles (fig. 6.10). Smaller sizes move more as a function of the *dS* product and seem to be relatively independent of velocity. Thus, the shear stress approach may be completely valid only for smaller sizes or low-velocity flows, and very fine-grained sediment requires higher velocities for its entrainment than the sixth-power law would predict.

In a related, but different type of approach, Bagnold (1973, 1977) proposed that entrainment and transportation of bedload can be analyzed in terms of stream power. **Stream power** is defined as

$$\omega = \gamma Q S$$

where $\omega$ is stream power and $\gamma$, $Q$, and $S$ are specific weight, discharge, and slope, respectively. If power is considered per unit area of the streambed, it essentially becomes a combination of shear stress and velocity because

$$\omega = \gamma QS/\text{width} = \gamma dSV = \tau V$$

where $V$ is mean velocity. Where available stream power is greater than that needed to transport load, scour of bed alluvium (entrainment) will occur. As a result, stream power has become an important parameter in characterizing the erosional capability of rivers.

Most of the studies of entrainment whether utilizing the shear stress, critical bed velocity, or stream power, have been based on flume studies. Flumes are not useful in the study of competence when particles are larger than pebble size. Most competence investigations of coarser sediment have, therefore, been made in natural rivers. These investigations demonstrate that in natural channels particles of a given size may be entrained by widely varying flow conditions. Much of this variation comes from the fact that the river bed is not composed of clasts of a uniform size, shape, and composition, but is a mixture of particles whose characteristics may vary over a considerable range. It is now known, for example, that in channels with poorly sorted bed material, the finer particles are shielded from the flow by the larger particles. Exposure of fine clasts may be particularly reduced by microtopographic features, such as pebble clusters (fig. 6.4) (Brayshaw 1985). The result of these "hiding effects" is that the larger clasts tend to be more mobile and the finer clasts less mobile than would be predicted for material of uniform size (Parker et al. 1982; Andrews 1983; Paola and Seal 1995). Entrainment may also be complicated in coarse-grained channels by the burial of a fine-particle layer by a coarser layer, a process that accentuates hiding effects and allows the larger particles to be more readily available for entrainment (Paola and Seal 1995). These observations suggest that the relative size of a particle in the mixture may be as important to entrainment as its absolute size. In fact, some investigators argue that the effects of particle-hiding and sediment layering may be so significant that all clasts in the mixture become mobile at about the same shear stress, a concept referred to as the *equal mobility hypothesis* (Parker et al. 1982; Andrews 1983; Andrews and Erman 1986).

A common phenomenon along many river systems is for particle size to decrease quasi-systematically downstream, although the reduction in size may be altered by the local influx of tributary sediment (Knighton 1980; Pizzuto 1995). Ashworth and Ferguson (1989)

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b. Grain-Size / Hydraulic Equations - What equations can one plug into?

| | |
|---|---|
| $T_c = 166\ d$ | |
| $D = 0.0001\ A^{1.21}\ S^{-0.57}$ | (Knox, 1987) |
| $V = 0.065\ d^{0.5}$ | (Williams, 1983) |
| $V_c = 0.18\ d^{0.44}$ | (Koster, 1978) |
| $V_c = 0.18\ d^{0.49}$ | (Costa, 1983) |
| $Q_{1.5} = 0.011\ L_m^{1.54}$ | (Williams, 1983) |
| $\lambda_m = 166\ Q_m^{0.46}$ | (Carlston, 1965) |
| $T = 0.030\ d^{1.49}$ | (Williams, 1983) |
| $T = 0.17\ d$ | (Williams, 1983) |
| $\omega = 0.079\ d^{1.29}$ | (Williams, 1983) |

Symbols (Williams, 1984)

$A$ = intermediate axis of largest clasts, mm
$d$ = particle diameter, mm
$D$ = competent flow depth, m
$\lambda_m$ = meander wavelength, m
$Q_{1.5}$ = discharge of 1.5 yr flood, $m^3/s$
$Q_m$ = mean annual discharge, $m^3/s$
$S$ = energy slope (approx. = topo. gradient), m/m
$V$ = mean flow velocity, m/s
$V_c$ = threshold (critical) flow velocity, m/s
$T_c$ = threshold (critical) tractive force, N/m
$T$ = bed shear stress, N/m
$\omega$ = stream power/m of width, watts/$m^2$

**Table 1.** Equations to yield entrainment threshold a-axis for boulders [from *Hopkins*, 1844]

| boulder cross section parallel to flow | equation from *Hopkins*@ | equation reduced$ solved for v |
|---|---|---|
| *terms*<br>a = streamwise axis of the boulder<br>a' = (spheroidal case) the average radius calculated as = $\sqrt{(0.5(a^2+c^2))}$ see note below<br>b = axis transverse or perpendicular to flow (all results are independent of this axis)<br>c = vertical axis<br>v = mean fluid velocity at threshold<br>μ = coefficient of friction, 1 = tan 45°<br>g = acceleration due to gravity<br>f = internal angle of the boulder<br>g = specific gravity of boulder - Hopkins used 2.5 when missing from equation below<br>g' = specific gravity of water (or the fluid in question), taken as 1<br>n = c/a, except for the spheroidal case when n = (a/(a-c)), where a>c<br>(note: Hopkins uses a and b for the two axes of the spheroid in cross section, but to allow for the b axis transverse to flow we have changed his b to c for consistency with the above. | | |
| *triangle* | $a = (1-(\mu/\sqrt{3}))(v^2/g)$ | $v = 4.82\sqrt{a}$ |
| for the equilateral - inside angle f = 60<br>when f represents the internal angle of the boulder, then he shows that for sliding to take place that it is required that tan $f > \mu$. He shows that no triangular section boulder can roll continuously. | | |
| *cube* | $a = (0.667n)(v^2/g)$ | $v = 3.84\sqrt{a}$ - sliding<br>[also found in analysis of *Graf*, 1979] |
| where n = c/a, and n = 1 represents a cubical section.<br>For c=na the boulder rolls for m >1/n, and when c=a it rolls if m > 1.<br>for rolling | | |
| | $a = (0.667/\mu))(v^2/g)$ | $v = 0.78\sqrt{a}$ - rolling |
| *pentagon*<br>almost identical equations for either sliding and rolling | $a = 0.568(v^2/g)$ | $v = 4.24\sqrt{a}$ |
| *hexagon*<br>almost identical equations for either sliding and rolling | $a= 0.57(v^2/g)$ | $v = 4.24\sqrt{a}$ |
| *spheroids** | $a = (n/6)(v^2/g)$<br>rolling | $v = 4.43\sqrt{a'}$<br>taking n = 3 |

@ we have organised the terms from *Hopkins* [1844]in a more consistent manner
$ we have taken g as 9.81m $s^{-2}$ and submerged specific gravity as 1.6 t $m^3$ rather than 1.5.
* for spheroids it is assumed that the *a* axis is not equal to the *b* axis, and the ratio n = (a/(a-c)) acts as an index of shape relative to a sphere.

same 'immovability'. For example, a boulder with a density of 2 g/$cm^3$ and length of 1 m, has the same $I_n$ value as a boulder with a density of 3 g/$cm^3$ and 0.5 m length. Interestingly, Butcher and Atkinson dismissed the importance of the coefficient of friction for sliding. They argued instead, that boulders are either stationary, or in unsteady motion owing to turbulence so that true sliding does not effectively occur.

For experimental verification of the theory, different sized boulders with the characteristic length measured in cm and with densities in the range 1.5 g/$cm^3$ to 4.0 g/$cm^3$ such that $I_{0.5}$, $I_1$ etc., were employed in flow models at three different scales to represent entrainment conditions on a full-scale concrete apron. Model results were consistent, demonstrating for example that for the field conditions the full scale value of *I* needed to be greater than 2, when units

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