1

Introduction to Hydrologic Science

1.1 DEFINITION AND SCOPE OF HYDROLOGY

Hydrology is broadly defined as the geoscience that describes and predicts the occurrence, circulation, and distribution of the water of the earth and its atmosphere. It has two principal focuses.

The global hydrologic cycle: the distribution and spatial and temporal variations of water substance in the terrestrial, oceanic, and atmospheric compartments of the global water system.

The land phase of the hydrologic cycle: the movement of water substance on and under the earth's land surfaces, the physical and chemical interactions with earth materials accompanying that movement, and the biological processes that conduct or affect that movement.[1]

Figure 1-1 shows the components of the global hydrologic cycle, and Figure 1-2 shows the storages and flows of energy and water that constitute the land phase of the cycle.

Horton (1931, p. 192) characterized the range of scales on which hydrologic processes operate:

> Any natural exposed surface may be considered as a unit area on which the hydrologic cycle operates. This includes, for example, an isolated tree, even a single leaf or twig of a growing plant, the roof of a building, the drainage basin of a river-system or any of its tributaries, an undrained glacial depression, a swamp, a glacier, a polar ice-cap, a group of sand dunes, a desert playa, a lake, an ocean, or the earth as a whole.

Figure 1-3 gives a quantitative sense of the range of time and space scales in the domain of hydrologic science.

Figure 1-4 shows the position of hydrologic science in the spectrum from basic sciences to water-resource management. Hydrology is built upon the basic sciences of mathematics, physics, chemistry, and biology, and it uses them as tools. It is an interdisciplinary geoscience, built also upon its sister geosciences, but differing from them in the subject matter on which it focuses. Much of the motivation for answering hydrologic questions has come, and will continue to come, from the practical need to manage water resources and water-related hazards. Thus, hydrologic science is (or should be) the basis for engineering hydrology and, along with economics and related social sciences, for water-resources management.

1.2 DEVELOPMENT OF SCIENTIFIC HYDROLOGY

Humans have been concerned with managing water as a necessity of life and as a potential hazard at least since the first civilizations developed along the banks of rivers. Hydraulic engineers built functioning canals, levees, dams, subterranean water con-

[1] This definition is consistent with that given by the U.S. Committee on Opportunities in the Hydrologic Sciences (Eagleson *et al.*, 1991, p. 57–58).

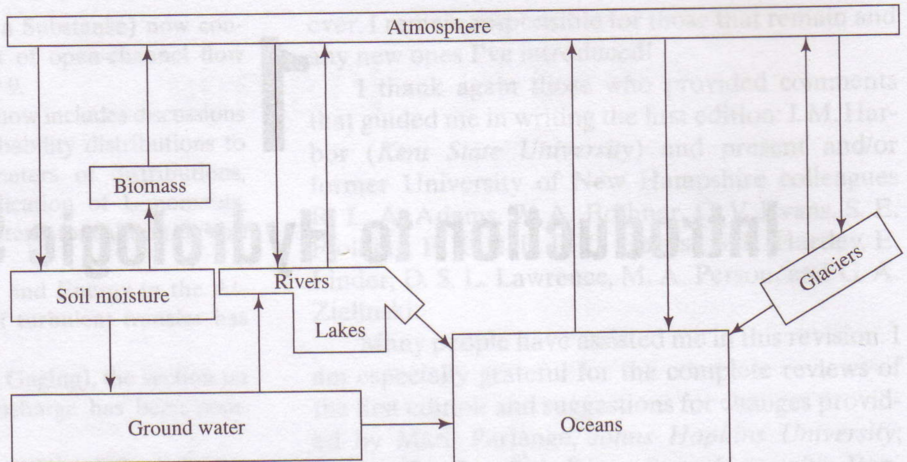


FIGURE 1-1 Principal storages (boxes) and pathways (arrows) of water in the global hydrologic cycle.

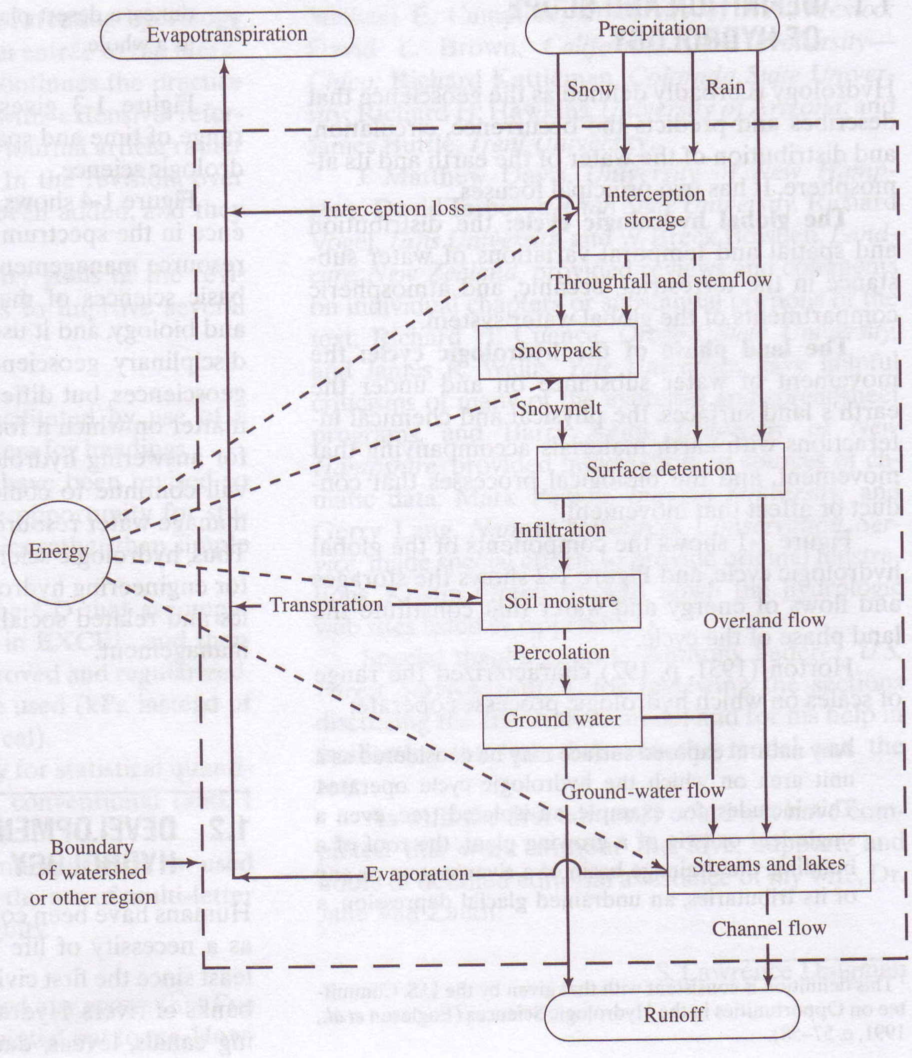


FIGURE 1-2 Principal storages (boxes) and pathways (arrows) of water in the land phase of the hydrologic cycle.

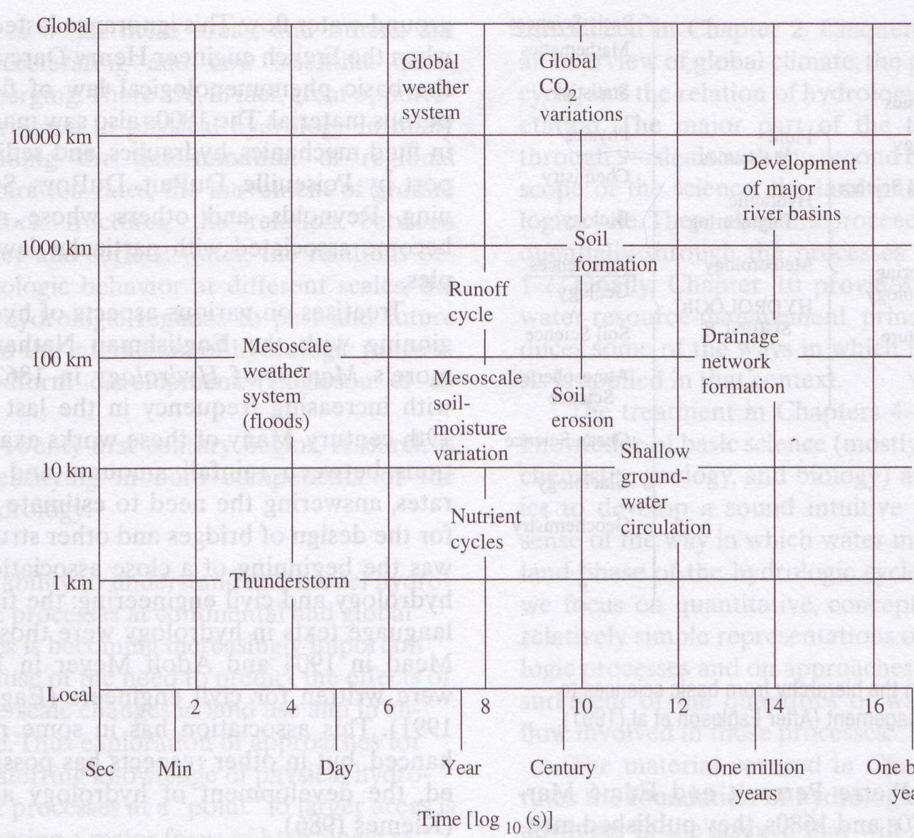


FIGURE 1-3
Range of space and time scales of hydrologic processes (After Eagleson et al (1991).

duits, and wells along the Indus in Pakistan, the Tigris and Euphrates in Mesopotamia, the Hwang Ho in China, and the Nile in Egypt as early as 5,000–6,000 years ago (B.P.). Hydroclimatologic information became vital to these civilizations; monitoring of river flows was begun by the Egyptians about 3,800 B.P., and the first known rainfall measurements are by Kautilya of India by 2,400 B.P. (Eagleson et al., 1991).

The concept of a global hydrologic cycle dates from at least 3,000 B.P. (Nace 1974), when Solomon wrote in Ecclesiastes 1:7 that

> All the rivers run into the sea; yet the sea is not full; unto the place from whence the rivers come, thither they return again.

Early Greek philosophers such as Thales, Anaxagoras, Herodotus, Hippocrates, Plato, and Aristotle also embraced the basic idea of the hydrologic cycle. However, while some of them had reasonable understandings of certain hydrologic processes, they postulated various fanciful underground mechanisms by which water returned from sea to land and entered rivers. The Romans had extensive practical knowledge of hydrology and, especially, hydraulics and developed extensive aqueduct systems; their "scientific" ideas, however, were based very closely on those of the Greeks.

The theories of the Greek philosophers continued to dominate western thought until the Renaissance, when Leonardo da Vinci (ca. 1500) in Italy and, most notably, Bernard Palissy (ca. 1550) in France asserted, on the basis of field observations, that the water in rivers comes from precipitation (Adams 1938; Biswas 1970). The modern scientific approach to the hydrologic cycle which they initiated was taken up in the 17th century by

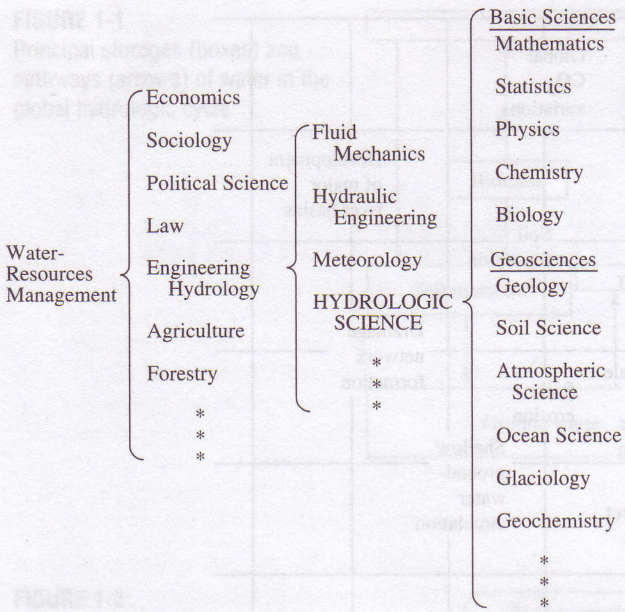


FIGURE 1-4
Hydrologic science in the hierarchy from basic sciences to water-resources management (After Eagleson et al (1991).

the Frenchmen Pierre Perrault and Edmé Marriotte: In the 1670s and 1680s, they published measurements and calculations that quantitatively verified the rainfall origin of streamflow. Shortly after (ca. 1700), the English scientist Edmund Halley extended the quantification of the hydrologic cycle by estimating the amounts of water involved in the ocean–atmosphere–rivers–ocean cycle of the Mediterranean Sea and its surrounding lands.

The 18th century saw considerable advances in applications of mathematics to fluid mechanics and hydraulics by Pitot, Bernoulli, Euler, Chézy, and others in Europe. Use of the term "hydrology" in approximately its current meaning began about 1750. By about 1800, the work of the English physicist and chemist John Dalton had firmly established the nature of evaporation and the present concepts of the global hydrologic cycle (Nace 1974), and Lyell, Hutton, and Playfair had published scientific work on the fluvial erosion of valleys. Routine network measurements of precipitation began before 1800 in Europe and the United States and were well established there and in India by 1820.

One of the barriers to understanding the hydrologic cycle was ignorance of the process of ground-water flow. This ignorance lasted until 1856, when the French engineer Henry Darcy established the basic phenomenological law of flow through porous material. The 1800s also saw many advances in fluid mechanics, hydraulics, and sediment transport by Poiseuille, DuPuit, DuBoys, Stokes, Manning, Reynolds, and others whose names have become associated with particular laws or principles.

Treatises on various aspects of hydrology, beginning with the Englishman Nathaniel Beardmore's *Manual of Hydrology* in 1862, appeared with increasing frequency in the last half of the 19th century. Many of these works examined relations between rainfall amounts and streamflow rates, answering the need to estimate flood flows for the design of bridges and other structures. This was the beginning of a close association between hydrology and civil engineering; the first English-language texts in hydrology were those of Daniel Mead in 1904 and Adolf Meyer in 1919, which were written for civil engineers (Eagleson et al. 1991). This association has in some respects enhanced, but in other respects has possibly inhibited, the development of hydrology as a science (Klemeš 1986).

The first half of the twentieth century saw great progress in many aspects of hydrology and, with the formation of the Section of Scientific Hydrology in the International Union of Geodesy and Geophysics (1922) and the Hydrology Section of the American Geophysical Union (1930), the first formal recognition of the scientific status of hydrology. Among those contributing notably to advances in particular areas in the early and middle decades of the century were the following: A. Hazen, E.J. Gumbel, H.E. Hurst, and W.B. Langbein in the application of statistics to hydrologic data; O.E. Meinzer, C.V. Theis, C.S. Slichter, and M.K. Hubbert in the development of the theoretical and practical aspects of ground-water hydraulics; L. Prandtl, T. Von Kármán, H. Rouse, V.T. Chow, G.K. Gilbert, and H.A. Einstein in stream hydraulics and sediment transport; R.E. Horton and L.B. Leopold in the understanding of runoff processes and quantitative geomorphology; W. Thornthwaite and H.E. Penman in the understanding of climatic aspects of hydrology and in the modeling of evapotranspiration; and A. Wolman and R.S. Garrels in the understanding and modeling of water quality.

Advances in all these areas, and others, are currently accelerating and new scientific questions are emerging. There are, in fact, great opportunities for progress in physical hydrology in many areas, including the determination of regional evapotranspiration rates, the movement of ground water in rock fractures, the relation between ground water and surface water, the relations between hydrologic behavior at different scales, the relation of hydrologic regimes to past and future climates, and the interaction of hydrologic processes and landform development (Eagleson et al. 1991).

As the twenty-first century begins, research is rapidly intensifying in both components of the scope of hydrology:

1. The ability to understand and model hydrologic processes at continental and global scales is becoming increasingly important because of the need to predict the effects of large-scale changes in land use and in climate. Thus exploration of approaches for transferring knowledge of physical hydrologic processes at a "point" to larger areas is becoming a major focus of hydrologic research, using newly available remote-sensing platforms and geographic-information systems.
2. In the land phase of the hydrologic cycle, it is interesting that detailed field studies to understand the mechanisms by which water enters streams began to proliferate only in the 1960s, pioneered by T. Dunne and others. Because of the temporal and spatial variability of natural conditions, this understanding is still far from complete. Research into these mechanisms is accelerating, motivated in part by concerns about the effects of land use on water quality and quantity and spurred by advances in technology that allow the use of a suite of chemical and isotopic tracers.

1.3 APPROACH AND SCOPE OF THIS BOOK

This text has four principal themes. First, the basic concepts underlying the science of hydrology are introduced in Chapter 2. Chapter 3 then provides an overview of global climate, the global hydrologic cycle, and the relation of hydrology to soils and vegetation. The major part of the text—Chapters 4 through 9—deals with the second component of the scope of the science: the land phase of the hydrologic cycle. These chapters proceed more or less sequentially through the processes shown in Figure 1-2. Finally, Chapter 10 provides an overview of water-resource-management principles and introduces some of the ways in which hydrologic analysis is applied in that context.

The treatment in Chapters 4–9 draws on your knowledge of basic science (mostly physics, but also chemistry, geology, and biology) and of mathematics to develop a sound intuitive and quantitative sense of the way in which water moves through the land phase of the hydrologic cycle. In this process, we focus on quantitative, conceptually sound, but relatively simple representations of physical hydrologic processes and on approaches to the field measurement of the quantities of water and rates of flow involved in those processes.

The material covered in Chapters 4–9 constitutes the foundation of hydrologic science, and the advances in the science that will come in the next decades—in understanding watershed response to rain and snowmelt, in forecasting the hydrologic effects of land use and climatic change over a range of spatial scales, in understanding and predicting water chemistry, and in other areas—will be built upon this foundation. However, we must be aware that an understanding of the basic physics of such processes as evaporation, snowmelt, and infiltration as they occur instantaneously at a given "point" (i.e., a small, relatively homogeneous region of the earth's surface) does not always extrapolate easily to an understanding of the hydrology of a finite area, such as a drainage basin, over a finite time. An important reason for this problem of scale is that hydrologic quantities and the factors that control them vary greatly in both space and time, and it is difficult and expensive to obtain data to characterize this variability.

Another reason for the difficulty in extrapolating from knowledge of processes at small space and time scales to larger scales is that, in general, rates of water movement are *nonlinear* functions of the controlling quantities. If q represents an instantaneous water-movement rate and x the instantaneous value(s) of the variable(s) that control that rate, we can often use physical principles to derive the functional relation between them:

$$q = f(x). \quad \textbf{(1-1)}$$

However, we are usually interested in the rate q averaged over a region (e.g., a watershed) or a period of time (e.g., a day) or both. If relation (1-1) is linear, we can measure spatial or temporal averages of x (denoted as $\overline{x}$) and compute the average flow rate ($\overline{q}$) as

$$\overline{q} = f(\overline{x}). \quad \textbf{(1-2)}$$

If, however, the relation is nonlinear,[2]

$$\overline{q} \neq f(\overline{x}). \quad \textbf{(1-3)}$$

This means that, even if we have information about $\overline{x}$ (and the use of satellite-based and other remotely-sensed information is an increasingly valuable source of such information) and good knowledge of $f(x)$, we might not be able to make good estimates of $\overline{q}$.

Thus, although the basic physical principles described in this text are powerful tools, the degree of knowledge that can be obtained with them is bounded, almost always by the limited spatial and temporal availability (and often the quality) of the data that characterize the field situation, and sometimes by the inherent problem of scaling as represented by Equation (1-3). Hydrologists must be as aware of these limitations as they are of the tools themselves. Thus, I have tried to point out the assumptions behind each conceptual approach and the difficulties in applying it, because

> It ain't so much the things we don't know that gets us in trouble. It's the things we know that ain't so.[3]

[2] You can easily demonstrate that relation (1-3) is true by some calculations with a simple nonlinear relation such as $q = x^2$ or $q = \ln(x)$.

[3] I have seen this quote attributed to three American humorists: Artemus Ward (pseudonym of Charles Farrar Browne), Mark Twain, and Will Rogers. Take your pick.

2

Basic Hydrologic Concepts

The concepts discussed in this chapter are so frequently applied in hydrology that they can be considered basic hydrologic concepts. Although hydrology is not a fundamental science in the sense that physics and chemistry are, its basic concepts are for the most part extensions of basic physical laws such as the conservation of mass.

2.1 PHYSICAL QUANTITIES AND LAWS

Hydrology is a quantitative geophysical science, and hydrologic relationships are usually expressed most usefully and concisely as relations between or among the numerical values of hydrologic quantities. In principle, these numerical values are determined by either

1. counting, in which case the quantity takes on a value that is a positive integer or zero; or
2. measuring, in which case the quantity takes on a value corresponding to a point on the real number scale that is the ratio of the magnitude of the quantity to the magnitude of a standard unit of measurement.[1]

Quantities determined by counting are **dimensionless** (dimensional quality expressed as [1]); measurable quantities have a dimensional quality expressed in terms of the fundamental physical dimensions force [F] (or mass [M]), length [L], time [T], and temperature [Θ]. Appendix A is a review of the rules for the treatment of dimensions and units in equations, for converting between different systems of units, and for handling significant figures.

The basic relations of physical hydrology are derived from fundamental laws of classical physics, particularly those listed in Table 2-1. Derivations begin with a statement of the appropriate fundamental law(s) in a mathematical form and with boundary and (if required) initial conditions appropriate to the situation under study and are carried out by using mathematical operations (algebra and calculus). This is the approach that we will usually follow in the discussions of hydrologic processes in this text.

The properties of water dictate how it responds to the forces that drive the hydrologic cycle; these properties are summarized in Appendix B.

2.2 HYDROLOGIC SYSTEMS

Several basic hydrologic concepts are related to the simple model of a **system** shown in Figure 2-1. For present purposes, a system is any conceptually defined region of space that is capable of receiving a sequence of **inputs** of a **conservative** quantity, storing some amount of that quantity, and discharg-

[1]Common temperature scales are *interval*, rather than ratio, scales and hence require designation of an arbitrary zero point as well as of a unit of measurement.

TABLE 2-1
Summary of Basic Laws of Classical Physics Most Often Applied in Hydrologic Analyses

Conservation of Mass

Mass is neither created nor destroyed.

Newton's Laws of Motion

1. The momentum of a body remains constant unless the body is acted upon by a net force (conservation of momentum).
2. The rate of change of momentum of a body is proportional to the net force acting on the body and is in the same direction as the net force. (Force equals mass times acceleration.)
3. For every net force acting on a body, there is a corresponding force of the same magnitude exerted by the body in the opposite direction.

Laws of Thermodynamics

1. Energy is neither created nor destroyed (conservation of energy).
2. No process is possible in which the sole result is the absorption of heat and its complete conversion into work.

Fick's First Law of Diffusion

A diffusing substance moves from where its concentration is larger to where its concentration is smaller at a rate that is proportional to the spatial gradient of concentration.

ing **outputs** of that quantity. The region is sometimes called the **control volume.** Note that a control volume can be defined to include regions that are not physically contiguous (e.g., the world's glaciers).

A conservative quantity is one that cannot be created or destroyed within the system. In the branch of physics known as mechanics, there are three conservative quantities: (1) mass ([M] or [F $L^{-1}T^2$]); (2) momentum ([M L T^{-1}] or [F T]); and (3) energy ([M L^2T^{-2}] or [F L]). In many hydrologic analyses, it is reasonable to assume that the mass density (mass per unit volume, [M L^{-3}]) of water is effectively constant; in these cases, volume [L^3] (i.e., [M] / [M L^{-3}]) is treated as a conservative quantity. However, mass density is a function of temperature (Section B.2.1), so this assumption might not always be warranted.

The storages and flows in Figures 1-1 and 1-2 are linked systems. The outer dashed line in Figure 1-2 indicates that any group of linked systems can be aggregated into a larger system; the smaller systems could then be called **subsystems.**

2.3 THE CONSERVATION EQUATIONS

The basic conservation equation can be stated in words as follows:

> The amount of a conservative quantity entering a control volume during a defined time period, minus the amount of the quantity leaving the volume during the time period, equals the change in the amount of the quantity stored in the volume during the time period.

Thus, the basic conservation equation is a generalization of the conservation of mass, Newton's first law of motion (when applied to momentum), and the first law of thermodynamics (when applied to energy). (See Table 2-1.)

In condensed form, we can state the conservation equation as

$$\text{Amount In} - \text{Amount Out} = \text{Change In Storage}, \qquad \textbf{(2–1)}$$

but we must remember that the equation is true *only* (1) for conservative substances, (2) for a defined control volume, and (3) for a defined time period.

If we designate the amount of a conservative quantity entering a region in time period Δt by I, the amount leaving during that period by Q, and the change in storage over that period as ΔS, we can write Equation (2–1) as

$$I - Q = \Delta S. \qquad \textbf{(2–2)}$$

Another useful form of the basic conservation equation can then be derived by dividing each of the terms in Equation (2–2) by Δt:

$$\frac{I}{\Delta t} - \frac{Q}{\Delta t} = \frac{\Delta S}{\Delta t}. \qquad \textbf{(2–3)}$$

If we now define the average rates of inflow, m_I, and outflow, m_Q, for the period Δt as follows:

$$m_I \equiv \frac{I}{\Delta t}, \qquad \textbf{(2–4)}$$

$$m_Q \equiv \frac{Q}{\Delta t}, \qquad \textbf{(2–5)}$$

we can write Equation (2–2) as

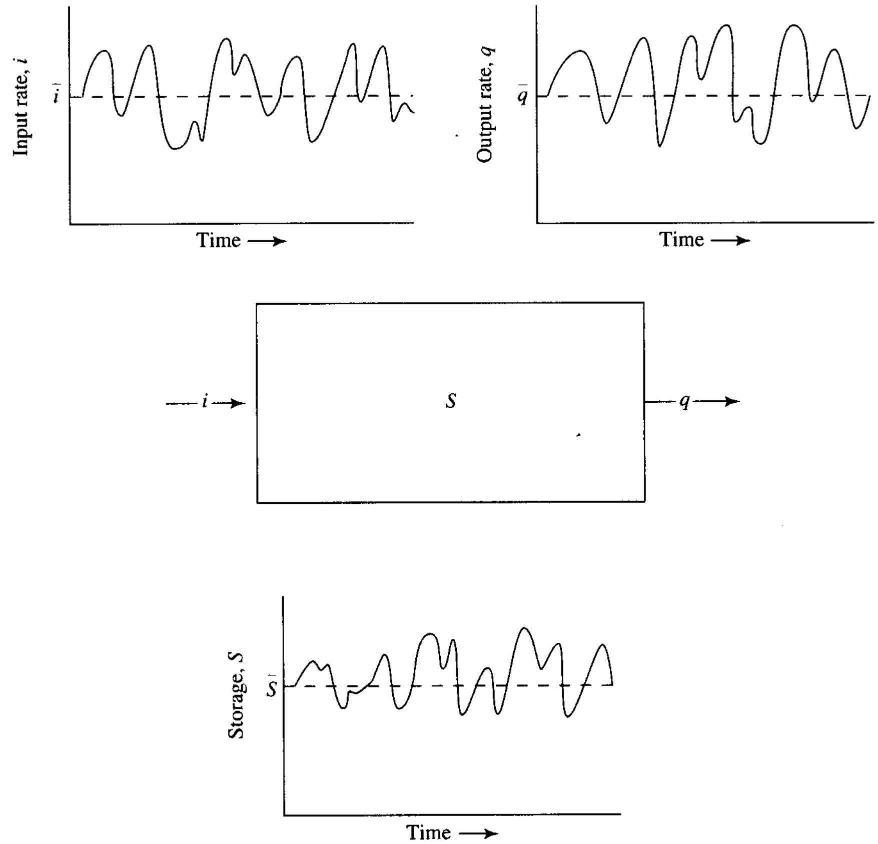


FIGURE 2-1
Conceptual diagram of a system. i is input rate, q is output rate, and S is storage.

$$m_I - m_Q = \frac{\Delta S}{\Delta t}. \tag{2–6}$$

Equation (2–6) states that the average rate of inflow, minus the average rate of outflow, equals the average rate of change of storage.

Another version of the conservation equation can be developed by defining the instantaneous rates of inflow, i, and outflow, q, as

$$i \equiv \lim_{\Delta t \to 0} \frac{I}{\Delta t} \tag{2–7}$$

and

$$q \equiv \lim_{\Delta t \to 0} \frac{Q}{\Delta t}, \tag{2–8}$$

respectively. Substituting these into Equation (2–3) allows us to write

$$i - q = \frac{dS}{dt}, \tag{2–9}$$

which states that the instantaneous rate of input, minus the instantaneous rate of output, equals the instantaneous rate of change of storage.

All three forms of the conservation equation, Equations (2–2), (2–6), and (2–9), are applied in various contexts throughout this text. They are called **water-balance equations** when applied to the mass of water moving through various portions of the hydrologic cycle; control volumes in these applications range in size from infinitesimal to global, and time intervals range from infinitesimal to annual or longer (Figure 1-3). A special application of these equations, the regional water balance, is discussed later in this chapter. As indicated in Figure 1-2, energy fluxes are directly involved in evaporation and snowmelt, and the application of the conservation equation in the form of **energy-balance equations** is essential to the understanding of those processes developed in Chapters 5 and 7. Consideration of the conservation of momentum is important in the analysis of fluid flow, and this principle is applied in the discussion of turbulent exchange of heat and water vapor with the atmosphere (Section D.6).

2.4 THE WATERSHED (DRAINAGE BASIN)

2.4.1 Definition

Hydrologists commonly apply the conservation equation in the form of a water-balance equation to a geographical region in order to establish the basic hydrologic characteristics of the region. Most commonly, the region is a **watershed** (also called **drainage basin, river basin,** or **catchment**), defined as the area that appears on the basis of topography to contribute all the water that passes through a given cross section of a stream (Figure 2-2). The surface trace of the boundary that delimits a watershed is called a **divide.** The horizontal projection of the area of a watershed is called the **drainage area** of the stream at (or above) the cross section.

The watershed concept is of fundamental importance because the water passing through the stream cross section at the watershed outlet originates as precipitation on the watershed[2] and because the characteristics of the watershed control the paths and rates of movement of water as it moves to the stream network. Hence, watershed geology, topography, and land cover determine the quality of ground water and of surface water as well as the magnitude and timing of streamflow and of ground-water outflow. As William Morris Davis stated in 1899,

> [O]ne may fairly extend the "river" all over its [watershed] and up to its very divides. Ordinarily treated, the river is like the veins of a leaf; broadly viewed it is like the entire leaf.

Thus the watershed can be viewed as a natural landscape unit, integrated by water flowing through the land phase of the hydrologic cycle and, although political boundaries do not generally follow watershed boundaries, water-resource and land-use planning agencies recognize that effective management of water quality and quantity requires a watershed perspective.

The location of the stream cross section that defines the watershed is determined by the purpose of the analysis. Hydrologists are most often interested in delineating watersheds above stream-gaging stations (where streamflow is measured; see Appendix F), or above points at which some water-resource activity takes place (e.g., a hydroelectric plant, a reservoir, a waste-discharge site, or a location where flood damages are of concern).

There are an infinite number of points (cross sections) along a stream, so an infinite number of watersheds can be drawn for any stream. As indicated in Figure 2-2, upstream watersheds are nested within, and are part of, downstream watersheds.

2.4.2 Delineation

The conventional manual method of watershed delineation requires a topographic map (or stereoscopically viewed aerial photographs). To trace the divide, start at the location of the chosen stream cross section, then draw a line away from the left or right bank, *maintaining it always at right angles to the contour lines.* Continue the line until it is generally above the headwaters of the stream network and its trend is generally opposite to the direction in which it began. Finally, return to the starting point and trace the divide from the other bank, eventually connecting it with the first line.

Frequent visual inspection of the contour pattern is required as the divide is traced out to assure that an imaginary drop of water falling streamward of the divide would, if the ground surface were imagined to be impermeable, flow downslope and eventually enter the stream network upstream of the starting point. A divide can never cross a stream, though there are rare cases where a divide cuts through a wetland (or, even more rarely, a lake) that has two outlets draining into separate stream systems. The lowest point in a drainage basin is always the basin outlet, i.e. (the starting point for the delineation). The highest point is usually, but not necessarily, on the divide.

Increasingly, topographic information is becoming available in the form of **digital elevation models** (DEMs). These are computer data files that give land-surface elevations at grid points. Although it is not an entirely straightforward exercise,

[2] As discussed in Section 2.5.2, there are situations in which ground-water inflow from adjacent watersheds contributes a portion (usually minor) of the flow of a stream.

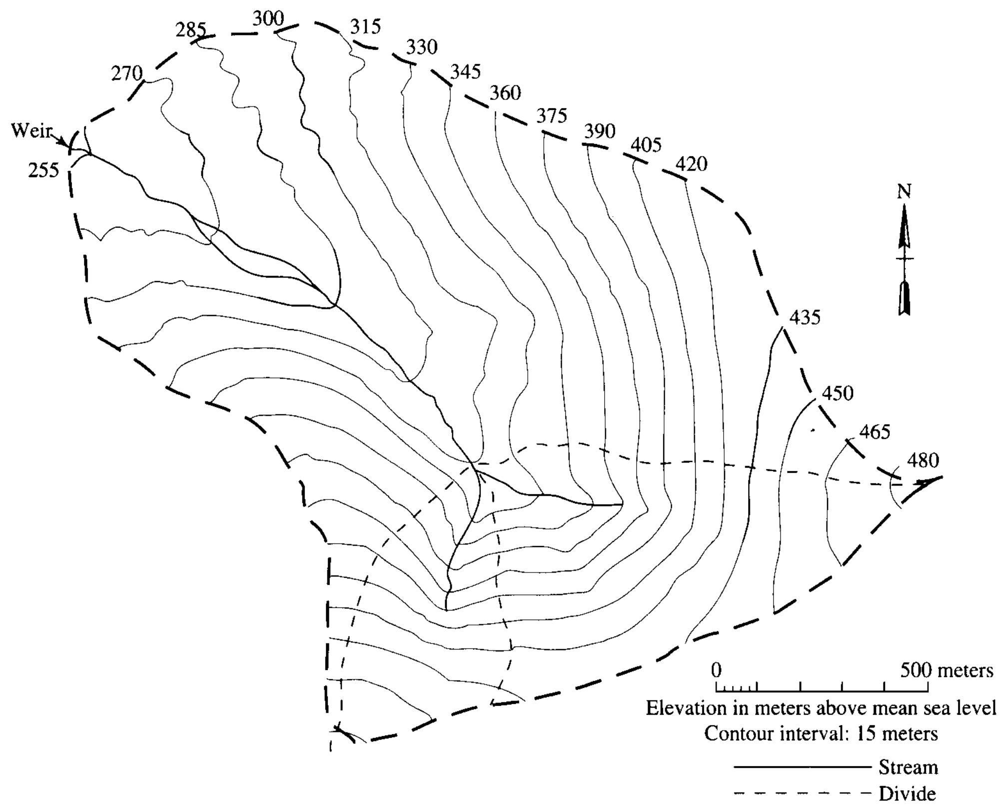


FIGURE 2-2
Watersheds delineated on a topographic map. Large area is the watershed of Glenn Creek in Fox, AK, above a streamflow-measurement weir; watersheds of tributaries to Glenn Creek are also shown.

it is possible to develop computer programs that can trace out stream networks and drainage divides by analyzing DEMs (e.g., Fairfield and Leymarie 1991; Martz and Garbrecht 1992; Tarboton 1997; McKay and Band 1998). This automated approach to watershed delineation allows the concomitant rapid extraction of much hydrologically useful information on watershed characteristics (such as the distribution of elevation and slope) that previously could be obtained only by very tedious manual methods.

2.5 THE REGIONAL WATER BALANCE

The regional water balance is the application of the water-balance equation to a watershed (or to any land area, such as a state or continent). Thus, the watershed area delimited by the divide (or other surface area) is the upper surface of the control volume; the sides of the volume extend vertically downward from the divide some indefinite distance that is assumed to reach below the level of significant ground-water movement.

In virtually all regional hydrologic analyses, it is reasonable to assume a constant density of water and to treat volume $[L^3]$ as a conservative quantity. For many such analyses, it is convenient to divide the volumes of water by the surface area of the region, so that the quantities have the dimension $[L](=[L^3]/[L^2])$; often, it is convenient also to divide by the duration of the measurement period, to obtain $[L\,T^{-1}]$. [Compare Equations (2-2) and (2-6).]

In this section, we first develop a conceptual regional water balance from which we can define some useful terms and show the importance of climate in determining regional water resources; we

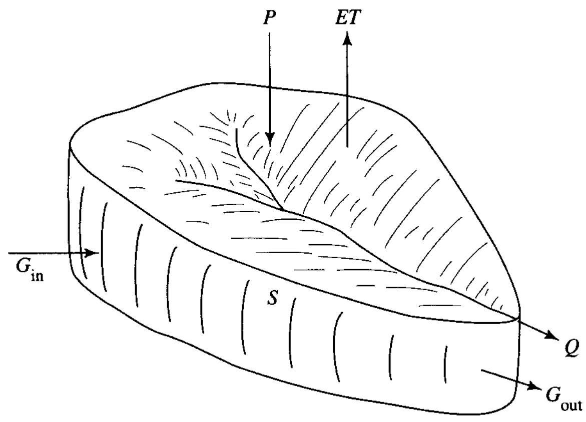


FIGURE 2-3
Schematic diagram of a watershed, showing the components of the regional water balance: P = precipitation, ET = evapotranspiration, Q = stream outflow, G_{in} = ground-water inflow, G_{out} = ground-water outflow.

then show how actual water-balance measurements are used to estimate regional evapotranspiration (defined in the next section).

2.5.1 The Water-Balance Equation

Consider the watershed shown in Figure 2-3. For any time period of length Δt, we can write the water-balance equation as

$$P + G_{in} - (Q + ET + G_{out}) = \Delta S, \qquad \textbf{(2–10)}$$

where P is precipitation (liquid and solid), G_{in} is ground-water inflow (liquid), Q is stream outflow (liquid), ET is evapotranspiration[3] (vapor), G_{out} is ground-water outflow (liquid), and ΔS is the change in all forms of storage (liquid and solid) over the time period. The dimensions of these quantities are $[L^3]$ (or, if divided by drainage area, [L]). If we average these quantities over a reasonably long time period (say, many years) in which there are no significant climatic trends or geological changes and no anthropogenic inputs, outputs, or storage modifications, we can usually assume that net changes in storage will be effectively zero and write the water balance as

$$\mu_P + \mu_{Gin} - [\mu_Q + \mu_{ET} + \mu_{Gout}] = 0, \qquad \textbf{(2–11)}$$

where μ denotes the time average of the subscript quantity and the dimensions are now $[L^3T^{-1}]$ or $[L\ T^{-1}]$.

The total amount of liquid water leaving the region is called the **runoff**,[4] RO, for the region. Therefore,

$$\mu_{RO} \equiv \mu_Q + \mu_{Gout}. \qquad \textbf{(2–12)}$$

The amount of liquid water actually "produced" in the region is called the **hydrologic production**, Π:

$$\mu_{\Pi} \equiv \mu_Q + \mu_{Gout} - \mu_{Gin} = \mu_P - \mu_{ET}. \qquad \textbf{(2–13)}$$

From Equations (2-11) and (2-12),

$$\mu_{RO} = \mu_P + \mu_{Gin} - \mu_{ET}. \qquad \textbf{(2–14)}$$

Because watersheds are defined topographically and ground-water flow is driven by gravity,[5] we can usually assume that G_{in} is negligible and write the water-balance equation as

$$\mu_{RO} \equiv \mu_Q + \mu_{Gout} = \mu_P - \mu_{ET}. \qquad \textbf{(2–15)}$$

Evaluation of the terms in the water-balance equation provides the most basic information about a region's hydrology. The runoff represents the water potentially available for human use and management and hence, the quantity of water resource available from a given region. However, as we will explore later in this chapter, the temporal variability of runoff must be evaluated in assessing actual water-resource availability.

As we will see in Chapter 7, evapotranspiration is determined largely (but not solely) by meteorologic variables (solar radiation, temperature, humidity, and wind speed), so both precipitation and evapotranspiration can be considered to be externally imposed climatic 'boundary conditions.' Thus,

[3]**Evapotranspiration** is the total of all water that leaves a region via direct evaporation from surface-water bodies, snow, and ice, plus that which is evaporated after passing through the vascular systems of plants (**transpiration**; the process is described in Chapter 7).

[4]Note that hydrologists also use the term "runoff" to denote overland flow, which is discussed in Chapter 9.

[5]The geometry of regional ground-water flows is extensively described in Chapter 8.

from Equation (2-15), runoff is a residual or difference between two climatically-determined quantities.

2.5.2 Estimation of Regional Evapotranspiration

Basic Approach

Perhaps the most common form of hydrologic analysis is the estimation of the long-term average value of regional evapotranspiration via the water-balance equation. This application arises because, although there are techniques for determining the precipitation over an area (discussed in Section 4.3) and the streamflow from an area (Appendix F), it is virtually impossible to measure areal evapotranspiration directly. (This is discussed in detail in Section 7.8.) In such analyses, it is usually assumed that ground-water flows either are negligible or cancel out and that ΔS is negligible, so that Equation (2-15) becomes

$$m_{ET} = m_P - m_Q, \qquad \textbf{(2–16)}$$

where m indicates the average of the subscript quantity for the period of measurement [rather than the true long-term averages as in Equation (2-15)].

Equation (2-16) is straightforward, but there are two types of errors that potentially affect the accuracy of m_{ET} estimates made with it: **model error**, which refers to the omission of potentially significant terms from the equation, and **measurement error** in the quantities m_P and m_Q, which is unavoidable. Together, these errors introduce uncertainty that propagates into the estimate of evapotranspiration and hence is critical for assessing the validity of that estimate. Thus, it is worthwhile to examine further these sources of error.

Model Error

Ground-Water Flows Ground-water flows are usually considered negligible in water-balance computations of regional evapotranspiration. However, the consideration of regional ground-water flows in Chapter 8 makes it clear that it is often unwise to assume that ground-water outflow is negligible. The higher the relief of a given watershed and the more hydraulically conductive its geologic composition (which is often not well known), the more likely it is to lose water by subsurface flow. Streams draining larger watersheds tend to receive the subsurface outflows of their smaller constituent watersheds, so the importance of ground-water outflow generally decreases as one considers larger and larger watersheds. However, such generalities can be obviated by particular geologic situations.

Storage Changes The net change in storage over a period of measurement is the difference between the amount of water in storage in the watershed (as ground water and as water in rivers, lakes, soil, vegetation, and snow and ice) at the end of the period and the amount in storage at the beginning of the period. No change-in-storage term appears in Equation (2-16), and this quantity is almost always assumed to be negligible.

Measurements of watershed storage are usually lacking, so the storage residual cannot be directly evaluated. Instead, hydrologists attempt to minimize its value by (1) using long measurement periods and (2) selecting the time of beginning and end of the measurement period such that storage values are likely to be nearly equal. (See Box 2-1.)

To minimize the storage residual in annual water-balance computations in the United States, the U.S. Geological Survey begins the **water year** on 1 October, on the assumption that by this time transpiration by plants will have largely ceased and soil-moisture and ground-water storage will have been recharged to near their maximum levels. However, as is suggested by the analysis in Box 2-1, other water-year spans may be more appropriate for specific regions—for example, the time of disappearance of the annual snowpack in northern areas.

Measurement Error

Uncertainty due to measurement error is always present in hydrologic computations, and sources of such error are reviewed in this text where various measurement techniques are discussed. Here we briefly characterize the uncertainty in estimating areal precipitation and streamflow and show quantitatively how this uncertainty is propagated into estimates of regional evapotranspiration when Equation (2-16) is assumed to be correct (i.e., when there is no model error).

Accuracy of Regional Precipitation Values In the application of Equation (2-16), it is assumed that m_P can

BOX 2-1

Evaluation of Changes in Watershed Storage

If S_i represents the watershed storage at the end of year i and ΔS_i the change in storage over year i, then the average change in storage over an N-year period, $m_{\Delta S}$, is

$$m_{\Delta S} = \frac{S_N - S_0}{N}. \quad \text{(2B1-1)}$$

Thus. $m_{\Delta S}$ will be small if $S_N - S_0$ is small or N is large. As noted in the text, $S_N - S_0$ can be minimized by choosing the beginning of the water year to be at a time when the year-to-year variability of storage is minimal.

Using the BROOK watershed model (Section 2.9.5) to simulate monthly evapotranspiration and soil-moisture storage over a 50-year period in New Hampshire, Hartley (1990) explored the consequences of using water years beginning on the first of each month. As shown in the table, she found a considerable variation in the statistics of annual storage changes depending on the month chosen as the beginning of the water year: choosing April gave the least year-to-year variability; choosing September gave the most. However, it turned out that the error in estimating average evapotranspiration introduced by assuming $\Delta S = 0$ was less than 5% after no more than two years regardless of the month selected.

Thus this study suggested that, although choosing different beginning times for water years leads to very different ΔS values, assuming $\Delta S = 0$ does not introduce significant error into water-balance estimates of evapotranspiration in this region if the averaging period exceeds a few years. However, ground-water and lake or wetland storage were not included in the simulation, and the conclusion might be different where these are important.

	Water Year Beginning 1st of											
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
$m_{\lvert\Delta S\rvert}$ (mm)	29.3	30.5	13.6	9.6	14.4	25.0	26.3	27.4	36.3	24.7	15.5	20.7
$s_{\Delta S}$ (mm)[a]	37.4	41.6	19.1	12.7	17.5	30.0	33.5	38.2	43.7	32.4	19.6	27.5
Maximum $\lvert\Delta S\rvert$ (mm)	93.3	117.2	62.8	32.9	42.2	62.3	96.3	106.8	77.9	78.0	52.4	81.9
Years for < 5% Error in m_{ET}	2	2	1	1	1	2	2	2	2	2	1	1

[a]Standard deviation of ΔS.

be calculated as the spatial average of temporally averaged precipitation-gage measurements made for many years in or near the region.

As discussed in Section 4.2.2, measurements of precipitation at individual gages are subject to error from several causes, and additional error is introduced in the process of computing areal averages. Thus, estimates of long-term (annual or longer) total (or average) areal precipitation typically have relative uncertainties on the order of 10% (i.e., the true value is taken to be within 10% of the estimated value) (Winter 1981). In regions of high relief or with few or poorly distributed gages, or for shorter measurement periods, the uncertainty can be considerably larger.

Accuracy of Streamflow Values In the application of Equation (2-16), it is assumed that m_Q can be calculated as the temporal average of streamflow measurements made at the watershed outlet for many years.

Winter (1981) estimated that the measurement uncertainty for long-term average values of streamflow at a gaging station is on the order of ±5%. (The accuracy of such measurements is discussed further in Section F.2.4.) Where μ_Q is estimated for locations other than carefully maintained gaging stations, the uncertainty can be much greater.

Propagation of Measurement Errors In general, both model and measurement error are present and are propagated into estimates of μ_{ET}. To simplify the discussion here, we show how uncertainty in the estimate of μ_{ET} via Equation (2-16) can be assessed quantitatively, given information about the uncertainties in the measurements of m_P and m_Q. To do this, we make the assumptions (1) that model errors in Equation (2-16) are negligible (i.e., that m_{Gin}, m_{Gout}, and the storage residual are negligible) and (2) that the terms in the equation refer to water-balance quantities for the period in which both P and Q were measured—that is, that we are not treating the data as samples from an indefinitely long time period.

Potential measurement errors are usually assumed to be distributed symmetrically about the true value (equal chance of under- or over-estimation) and to follow the bell-shaped **normal distribution** described in Appendix C: the further a measured value is from the true value (i.e., the larger is the error), the smaller is the probability that it will occur (Figure 2-4). The spread, or variation, of the potential measured values about the true value is expressed as the **standard deviation** of the potential errors.

We can apply to Equation (2-16) the general rule that, if a quantity (m_{ET}) is the sum or difference of two measured quantities (m_P and m_Q), and the errors in the two terms are normally distributed, then the errors in m_{ET} are also normally distributed. Furthermore, if the measurement errors of P and Q are not related (a reasonable assumption), then the standard deviations of the errors due to measurement of the quantities are related as

$$\sigma_{ET}^2 = \sigma_P^2 + \sigma_Q^2, \qquad \textbf{(2–17)}$$

where σ indicates the standard deviation of the measurement errors of the subscripted quantity.[6] Thus, to evaluate the error in ET due to measurement errors in P and Q via Equation (2-17), we must estimate σ_P and σ_Q.

Statements of measurement uncertainty are usually expressed probabilistically—for example, as

"I am $100 \cdot p$% sure that the true value of precipitation is within $u_P \cdot m_P$ of the measured value." **(2–18a)**

[6]Equation (2-17) can be derived directly from the definition of the standard deviation [Equation (C-19)].

Here, m_P is the estimate of average precipitation and u_P is the **relative uncertainty** in the estimate (e.g., if the measurement uncertainty is stated to be 10%, $u_P = 0.1$). The **absolute uncertainty** in m_P is $u_P \cdot m_P$. In conventional probability notation, Equation (2-18a) is written as

$$\Pr\{(m_P - u_P \cdot m_P) \leq \text{true precipitation} \leq (m_P + u_P \cdot m_P)\} = p, \qquad \textbf{(2–18b)}$$

where Pr{ } indicates the probability of the statement in braces.

The probability p should be close to 1, but it is not realistic to assume that $p = 1$; that would be equivalent to stating, "I am absolutely certain that the true value is within $\pm u_P \cdot m_P \ldots$". Typically, statements of measurement error are given with $p = 0.95$, so the statement would be "I am 95% sure. . . ."

Given that potential measurement errors follow the normal distribution, we can find from the properties of that distribution, summarized in Table C-5, that there is a 95% probability that an observation will be within ±1.96 standard deviations of the central (true) value. Thus we can write Equation (2-18b) equivalently as

$$\Pr\{(m_P - 1.96 \cdot s_P) \leq \text{true precipitation} \leq (m_P + 1.96 \cdot s_P)\} = 0.95, \qquad \textbf{(2–19)}$$

where s_P is the estimated error standard deviation for precipitation.

From comparison of Equations (2-18) and (2-19), we see that when $p = 0.95$,

$$u_P \cdot m_P = 1.96 \cdot s_P, \qquad \textbf{(2–20)}$$

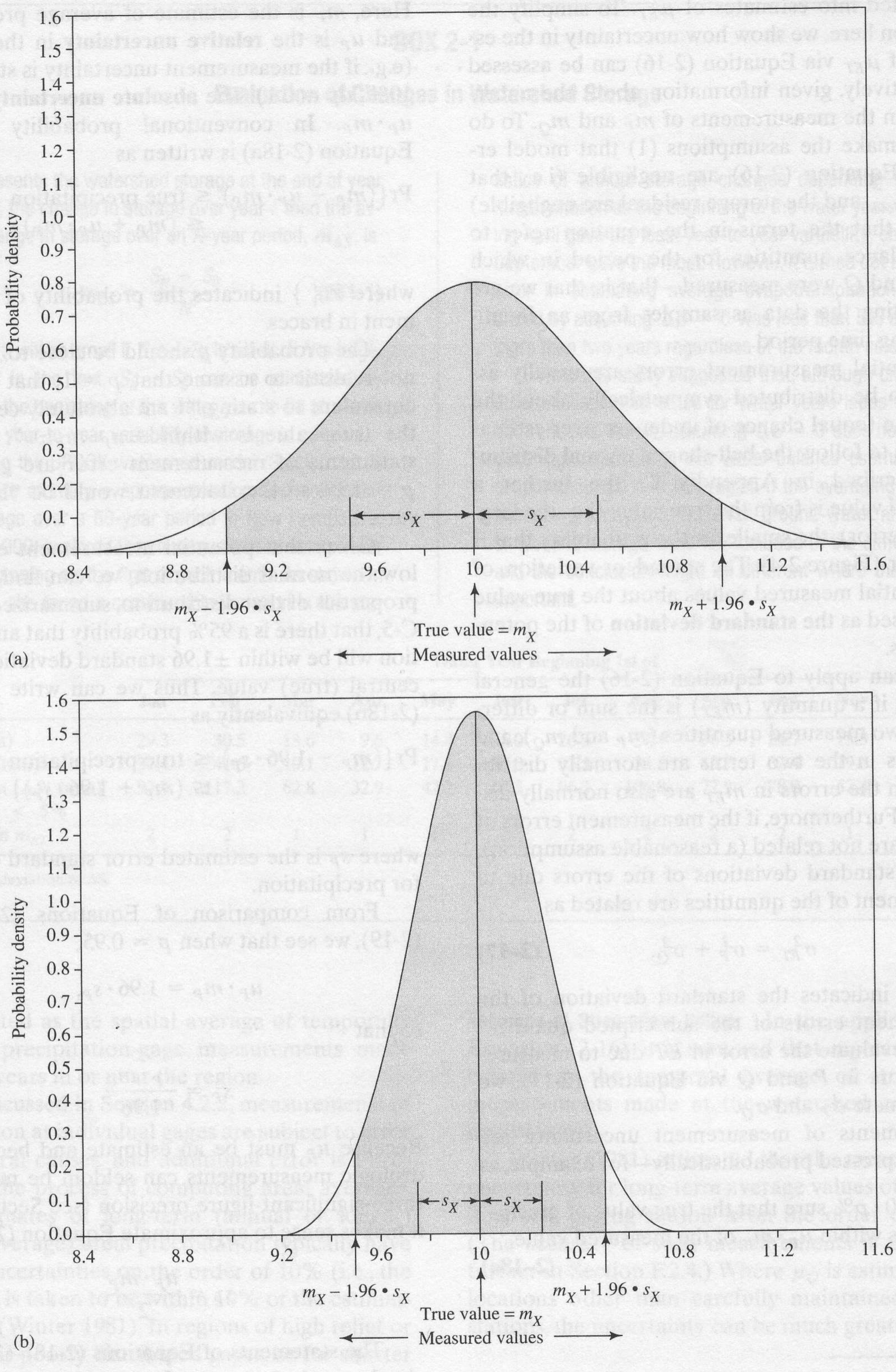
so that

$$s_P = \frac{u_P \cdot m_P}{1.96}. \qquad \textbf{(2–21a)}$$

Because u_P must be an estimate and because hydrologic measurements can seldom be made with three-significant-figure precision (see Section A.3), it makes sense to approximate Equation (2-21a) as

$$s_P = \frac{u_P \cdot m_P}{2}. \qquad \textbf{(2–21b)}$$

The statements of Equations (2-18)–(2-21) can also be made for streamflow and its measurement errors, so we can also conclude that


Probability density
1.6
1.5
1.4
1.3
1.2
1.1
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
8.4
8.8
9.2
9.6
10
10.4
10.8
11.2
11.6
s_X
s_X
$m_X - 1.96 \cdot s_X$
$m_X + 1.96 \cdot s_X$
True value = m_X
Measured values
(a)
Probability density
1.6
1.5
1.4
1.3
1.2
1.1
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
8.4
8.8
9.2
9.6
10
10.4
10.8
11.2
11.6
s_X
s_X
$m_X - 1.96 \cdot s_X$
$m_X + 1.96 \cdot s_X$
True value = m_X
Measured values
(b)

$$s_Q = \frac{u_Q \cdot m_Q}{2}, \qquad (2\text{–}22)$$

where u_Q is the relative error in streamflow measurement (and $p = 0.95$).

Thus, if reasonable estimates of u_P and u_Q can be obtained, s_{ET} is readily calculated via Equation (2-17). By analogy with Equations (2-21) and (2-22), the relative error in the evapotranspiration estimate, u_{ET}, is then

$$u_{ET} = \frac{2 \cdot s_{ET}}{m_{ET}} = \frac{(u_P^2 \cdot m_P^2 + u_Q^2 \cdot m_Q^2)^{1/2}}{m_P - m_Q}. \qquad (2\text{–}23)$$

EXAMPLE 2-1

For the period 1961–1985, average annual precipitation for the Oyster River drainage basin, NH, was 1066 mm yr^{-1}, and average annual streamflow was 551 mm yr^{-1}. Assume the relative measurement errors for precipitation and streamflow are $u_P = 0.1$ and $u_Q = 0.05$. Estimate (a) the average annual evapotranspiration for that period and (b) the relative and absolute uncertainties in that estimate.

Solution: (a) With $m_P = 1066$ mm yr^{-1} and $m_Q = 551$ mm yr^{-1}, and assuming that m_{Gin}, m_{Gout}, and the storage residual are all negligible, Equation (2-16) yields

$$m_{ET} = 1066 \text{ mm yr}^{-1} - 551 \text{ mm yr}^{-1} = 515 \text{ mm yr}^{-1}.$$

(b) From Equations (2-21b) and (2-22),

$$s_P = \frac{0.1 \times 1066 \text{ mm yr}^{-1}}{2} = 53.3 \text{ mm yr}^{-1};$$

$$s_Q = \frac{0.05 \times 551 \text{ mm yr}^{-1}}{2} = 13.8 \text{ mm yr}^{-1}.$$

Using these values in Equation (2-17) gives

$$s_{ET}^2 = (53.3 \text{ mm yr}^{-1})^2 + (13.8 \text{ mm yr}^{-1})^2$$
$$= 3031.33 \text{ mm}^2 \text{ yr}^{-2};$$
$$s_{ET} = 55.1 \text{ mm yr}^{-1}.$$

Thus, from Equation (2-23),

$$u_{ET} = \frac{2 \times 55.1 \text{ mm yr}^{-1}}{515 \text{ mm yr}^{-1}} = 0.214.$$

By analogy with Equation (2-19),

$$\Pr\{(515 \text{ mm yr}^{-1} - 0.214 \times 515 \text{ mm yr}^{-1}) \le \mu_{ET}$$
$$\le (515 \text{ mm yr}^{-1} + 0.214 \times 515 \text{ mm yr}^{-1})\} = 0.95;$$

$$\Pr\{405 \text{ mm yr}^{-1} \le \mu_{ET} \le 625 \text{ mm yr}^{-1}\} = 0.95.$$

The result of Example 2-1 is quite general: the relative uncertainty in estimates of μ_{ET} found via regional water balances is usually considerably greater than the uncertainties in the measured quantities, even when there are no unmeasured water-balance components and when storage residuals are negligible. More generally, the uncertainty in *any* quantity found as the difference of measured quantities is larger than the uncertainties in the measured quantities.

The studies by Lesack (1993) and Cook et al. (1998) are among the few published attempts to assess uncertainty in regional water balances.

FIGURE 2-4
Probability distribution of potential measurement errors of a quantity X, shown as having a true value $m_X = 10$. Such errors are usually assumed to be symmetrical about the true value and to follow the normal distribution (see Appendix C). (a) shows the case where one is 95% sure that the true value is within 10% of the measured value, so that $s_X = 0.1 \cdot m_X/1.96 \approx 0.5$. (b) Shows the case where one is 95 % sure the true value is within 5 % of the measured value, so that $s_X = 0.05 \cdot m_X/1.96 \approx 0.25$. In both cases the shaded area = 0.95 (the probability p).

2.6 SPATIAL VARIABILITY

Rates of input and output—and many other hydrologically relevant properties—vary spatially over the geographic regions that constitute control volumes for many types of hydrologic analyses (e.g., watersheds). Thus it is essential that hydrologists become familiar with methods for describing and comparing spatial as well as temporal variability.

Descriptions of spatial variability of some variables—notably precipitation—are based on measurements made over time at discrete points (precipitation gages). These measurements, expressed as average rates or total amounts, constitute spatial as well as temporal samples, and the values can be contoured to produce a model of the quantity's continuous spatial variation.

Traditional statistical methods, such as those described in Appendix C, can also be used to compute spatial averages and measures of spatial variability from the point values. However, precipitation gages are usually unevenly distributed over any given region, and the point values are therefore an unrepresentative sample of the true precipitation field. Because of this, and because of the importance of accurately quantifying variables such as precipitation, basic statistical concepts have been incorporated into special techniques for characterizing and accounting for spatial variability. These techniques are introduced in Section 4.3; however, they apply not only to rainfall, but to the spatial variability of infiltration, of ground-water levels, and of other spatially distributed quantities as well.

2.7 TEMPORAL VARIABILITY

The inputs, storages, and outputs in Figures 1-1, 1-2, and 2-1 are all **time-distributed variables**—quantities that can vary with time. Thus the concept of time variability is inherent to the concept of the system, and we have seen how time averaging is applied to develop alternative forms of the conservation equations.

In particular, the streamflow rate at a given location is highly variable in time. Even in humid regions, it typically varies annually over three or more orders of magnitude as a result of seasonal fluctuations of rainfall and evapotranspiration; in arid regions, the annual fluctuations are even greater. Year-to-year weather variations cause further temporal variability, reflected in fluctuations in annual mean flows and in the occurrences of floods and droughts.

From the human viewpoint, the long-term average streamflow rate, μ_Q, is highly significant: it represents the maximum rate at which water is potentially available for human use and management, and is therefore a measure of the ultimate water resources of a watershed or region. However, because of the large time variability of streamflow, we generally cannot rely on the mean flow to be available most of the time. The rate at which water is *actually* available for use is best measured as the streamflow rate that is available a large percentage—say 95%—of the time. This value is designated $q_{.95}$. Where streamflow records are available, $q_{.95}$ is readily determined by constructing a flow-duration curve, as described in Sections 2.7.2 and 10.2.5.

Streamflow variability is directly related to the seasonal and interannual variability of runoff (and hence of the climate of precipitation and evapotranspiration) and inversely to the amount of storage in the watershed. Humans can increase water availability by building storage reservoirs, as discussed in Sections 2.8 and 10.2.5. Humans can also attempt to increase μ_P through "rain-making" (Section 4.4.5) and to decrease μ_{ET} by modifying vegetation (Sections 7.6.4 and 10.2.5). However, such interferences in the natural hydrologic cycle usually have serious environmental, social, economic, and legal consequences. Some of the consequences involved in exploiting ground water are considered in Sections 8.6 and 10.2.4.

Because streamflow is the difference between two climatically determined quantities [Equation (2-15)], it is clear that climate change, whether natural or anthropogenic, will affect runoff and hence water resources. The BROOK90 model introduced in Section 2.9.5 can be used for detailed study of climatic and land-use effects; some simpler approaches to evaluating these effects are given in Chapter 3 (Boxes 3-4 and 3-5).

In the next section, we introduce (1) the basic approach for constructing and analyzing samples of time-distributed variables and (2) the duration curve, a widely applicable approach to characterizing time variability.

2.7.1 Time Series

Clearly, it will be useful to be able to describe and compare time-distributed variables in terms of their average value, variability, and perhaps other characteristics. Such descriptions and comparisons usually are made by applying the statistical methods described in Appendix C. However, these methods are applicable only to discrete sequences of values obtained from the time trace of the variable of interest, each value of which is associated with a particular time in a sequence of times. Such a sequence is called a **time series.**

Some time-series variables are obtained by counting—for example, the number of days with more than 1 mm rain in each year at a particular location. Such variables are inherently **discrete**, and it is a straightforward matter to construct a time series for such variables if the relevant measurements are available.

However, many hydrologic variables—including the inputs, outputs, and storages in Figure 2-1—are **continuous** time traces: they take on values at every instant in time. Thus there are infinitely many values of the variable associated with each time interval. In order to construct a time series for a continuous variable, one must convert it to discrete form. To do this, first select a time interval, Δt, and divide the total period of interest into increments of length Δt. (The value of Δt is determined by the purpose of the analysis; in hydrologic studies it is often 1 day, 1 month, or 1 year.) Then, the single value of the variable of interest associated with each interval is determined either as (1) the average, (2) the largest, or (3) the smallest value of the variable that occurred during the interval (for inputs and outputs) or as (4) the value of the variable at the beginning or end of each Δt (for storages). (See Figure 2-5.) Example 2-2 and Figure 2-6 show examples of three time series developed from the continuous measurements at a streamflow gaging station.

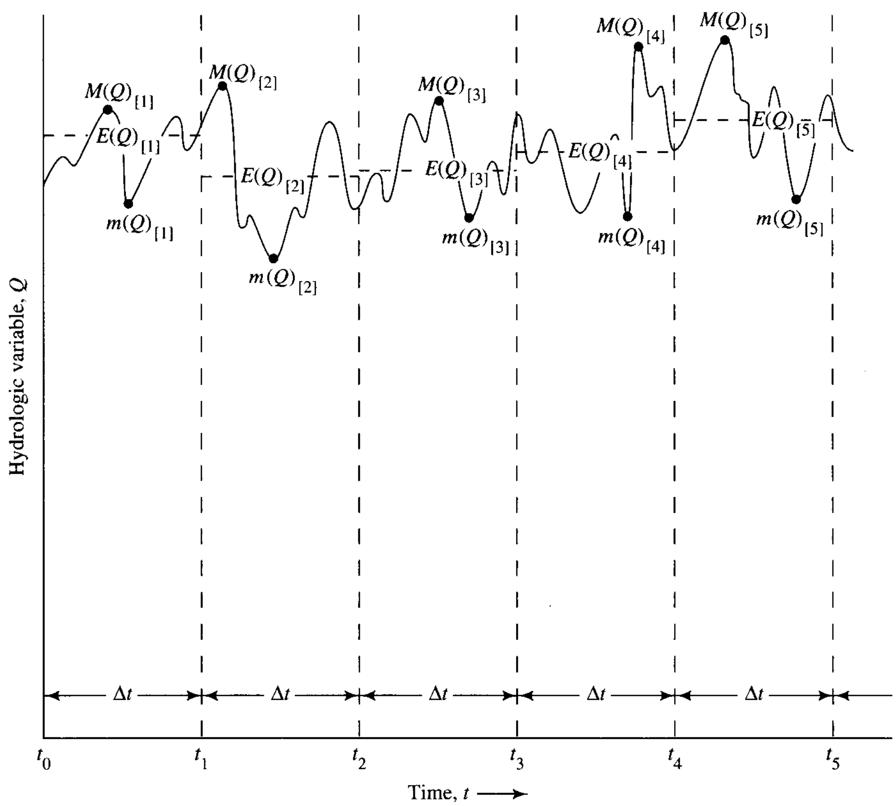


FIGURE 2-5
Schemes for converting a continuous time trace into a discrete time series. For each Δt, one may select the average, $E(Q)$, maximum, $M(Q)$, or minimum, $m(Q)$.

EXAMPLE 2-2

Table 2-2 lists, and Figure 2-6 plots, three time series developed from the continuous streamflow record obtained at the **stream-gaging station** operated by the U.S. Geological Survey on the Oyster River in Durham, NH. In all three plots, $\Delta t = 1$ yr, and the ordinate is a streamflow rate, or **discharge**, $[L^3T^{-1}]$. However, the discretization of the continuous record was done differently for each series: Series *a* is the average streamflow for the year, Series *b* is the highest instantaneous flow rate for the year, and Series *c* is the lowest of the flow rates found by averaging over all the seven-consecutive-day periods within each year. (These data are also in the spreadsheet file `Table2-2.xls` on the disk accompanying this text.)

Note that the lines connecting the time-series values in each graph do not represent a time trace; they serve only to

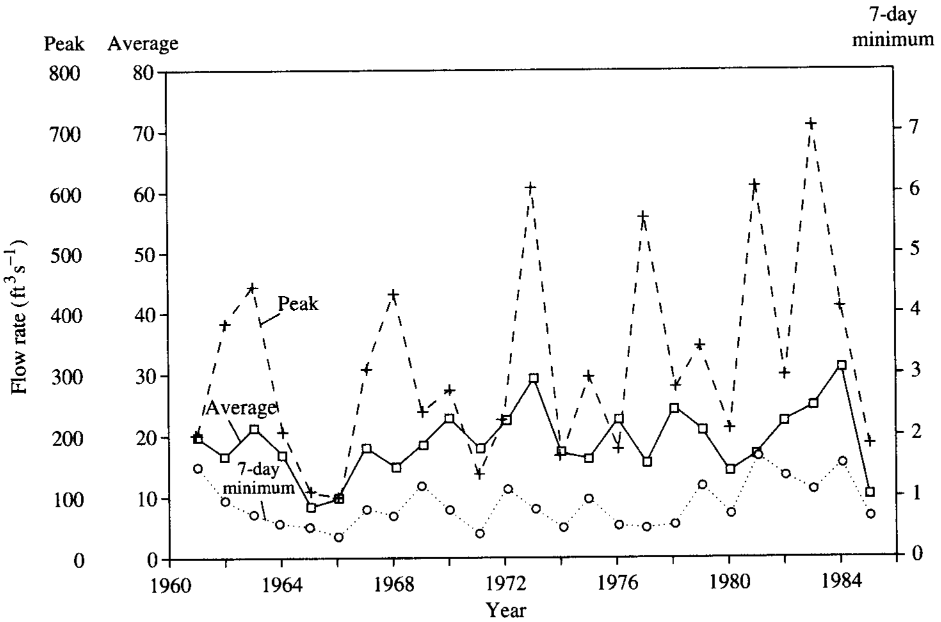


FIGURE 2-6
Plots of the time series in Table 2-2.

connect the point values to provide a visual impression of the nature of the series.

Time series are usually treated as more or less representative **samples** of the long-term behavior of the variable and are described and compared on the basis of their statistical attributes. For example, the temporal variability of a time series can be characterized in absolute terms by its **interquartile range** or by its **standard deviation**, and in relative terms by the ratio of its interquartile range to its **median** or by its **coefficient of variation** (Sections C.2.4 and C.2.5).

It is important to note that time series developed from a single continuous time trace by choosing different discretizing schemes (as in Example 2-2) or different Δt values will in general have very different statistical characteristics. Box 2-2 explains how one can obtain time series of peak and daily average streamflows measured at U.S. Geological Survey gaging stations.

2.7.2 Duration Curves

One conceptually simple but highly informative way to summarize the variability of a time series is by means of a **duration curve**—a cumulative-frequency curve that shows the fraction (percent) of time that the magnitude of a given variable exceeds a specified value over a period of observation that is long enough to include a wide range of seasonal and inter-annual variability. Duration curves are most commonly used to depict the temporal variability of streamflow; such curves are then called **flow-duration curves** (FDCs) (Figure 2-7).

Searcy (1959) and Vogel and Fennessey (1994; 1995) have provided comprehensive reviews of FDCs. Here we examine the general characteristics and interpretation of FDCs; their construction for stream locations with and without long-term streamflow records is discussed in Section 10.2.5. FDCs can be developed for flows averaged over periods (Δt) of any length—days, months, or years. However, we will focus exclusively on FDCs of daily average flows (Δt = 1 day), which are by far the most commonly used.

Statistical Interpretation

In statistical terms, the FDC is a graph plotting the magnitudes, q, of the variable Q (average daily flow, y-axis) vs. the fraction of time, $EP_Q(q)$, that Q exceeds any specified value $Q = q$ (x-axis). $EP_Q(q)$ is called the **exceedence probability** (or **exceedence**

TABLE 2-2
Three Time Series Developed by Discretization of the Continuous Streamflow Record for the Oyster River for the Period 1961–1985[a]

Year	Annual Average	Peak	Annual Seven-Day Minimum
1961	20.5	213	1.56
1962	17.5	386	0.97
1963	22.0	450	0.76
1964	17.5	213	0.61
1965	9.1	110	0.55
1966	10.2	106	0.43
1967	18.8	309	0.86
1968	15.3	440	0.74
1969	18.9	240	1.26
1970	23.3	280	0.79
1971	18.0	140	0.43
1972	22.5	233	1.17
1973	29.7	610	0.83
1974	17.8	169	0.52
1975	16.8	299	1.01
1976	23.2	179	0.55
1977	15.6	567	0.52
1978	24.9	278	0.56
1979	21.2	348	1.19
1980	14.1	210	0.73
1981	17.3	615	1.64
1982	22.5	287	1.34
1983	25.0	709	1.13
1984	31.5	410	1.57
1985	10.3	190	0.68

[a]All streamflows are in ft^3s^{-1}. See Example 2-2 for explanation.

frequency), and it is related to the pth quantile of streamflow, q_p, as

$$EP_Q(q_p) = 1 - F_Q(q_p) = \Pr\{Q > q_p\}, \quad (2\text{–}24)$$

where F_Q designates the cumulative distribution function (Section C.2) of Q and $\Pr\{\ \}$ denotes the probability of the condition within the braces.

For FDCs, exceedence frequency refers to the frequency or probability of exceedence in a "suitably long" period rather than the probability of exceedence on any specific day. Seasonal effects and hydrological persistence (Section C.9) cause exceedence probabilities of daily flows to vary as a function of time of year and antecedent conditions; the FDC does not account for those dependencies. By contrast, exceedence probabilities for flood flows and low flows are usually calculated on an annual basis ($\Delta t = 1$ yr) and do not vary from year to year.

It can be shown that the integral of the FDC is equal to the average flow for the period plotted; if that period is the period of measurement, the integral equals m_Q in the regional water balance [Equations (2-11)–(2-16)][7]. The flow exceeded on 50% of the days, $q_{.50}$, is the median flow. The distribution of daily streamflows is almost always highly skewed, so the mean flow is much larger than the median flow—in many humid regions, the mean flow is exceeded on only 20 to 30% of the days. The steepness of the FDC is proportional to the variability of the daily flows.

Comparison and Controlling Factors

For comparing flow characteristics among streams, it is often useful to plot q/A, (where A is drainage area) or q/m_Q on the vertical axis of the FDC. Using either q/A or q/m_Q eliminates the effect of drainage-basin size; using q/m_Q also eliminates the effect of differing per-unit-area precipitation and/or evapotranspiration rates. Either approach can reveal similarities in FDCs for streams in a region that might be useful in inferring the FDC for streams lacking long-term flow records (Box 10-3).

For streams unaffected by diversion, regulation, or land-use modification, the slope of the upper end of the FDC is determined principally by the regional climate, and the slope of the lower end by the geology and topography. The slope of the upper end of the FDC is usually relatively flat where snowmelt is a principal cause of floods and for large streams where floods are caused by storms that last several days. "Flashy" streams, where floods are typically generated by intense storms of short duration, have steep upper-end slopes. At the lower end of the FDC, a flat slope usually indicates that flows come from significant storage in ground-water aquifers (Section 8.5.3) or in large lakes or wetlands; a steep slope indicates an absence of significant storage. However, Dingman (1978b, 1981a) found that more frequent inputs of precipitation can also produce relatively flat slopes in the low ends of FDCs. The effect of reser-

[7]Graphically, the area under the FDC is proportional to m_Q if both axes have arithmetic scales.

BOX 2-2

Obtaining U.S. Geological Survey Streamflow Data in Spreadsheet Format

The Water Resources Division (WRD) of the U.S. Geological Survey (USGS) maintains over 5000 stream gages throughout the United States and publishes data on peak flows (floods), daily average flows, and water-quality parameters measured at these gages. The data are published annually in books entitled "Water Resources of [state(s)]" that are available from district offices of the WRD and electronically via the Internet. Spreadsheet files of these data can readily be obtained by downloading the data by using the following steps:

1. Create a directory, or insert a formatted floppy disk in your computer to receive the data.
2. Use your Internet browser to access the address

   http://h2o.usgs.gov/nwis-w/US

   to go to the national data base, from which you can select the state of interest; or replace the "US" with a state abbreviation to go directly to data files for gages in that state.
3. Click your mouse on the appropriate area to access historical streamflow data.
4. Select the gaging station of interest from a map, a statewide list, or lists organized by county or river basin.
5. Select data on peak flows or daily average flows. For peak flows, you can choose between (a) all **peaks above a base value** that the USGS has determined for the gage (there could be more than one peak for each water year, or none); and (b) **annual peaks** (that is, the highest peak flow in each water year).
6. The entire available period of record is displayed; you can select it or change the beginning or ending dates of interest.
7. For spreadsheet downloading, select "tab-delimited" rather than "punchcard image" format.
8. Select a date format.
9. Select "Retrieve Data". An image of the header information and data will appear on your screen.
10. Select "Save As" from the "File" menu of your Program Manager menu bar, designate the directory and file destination, and click "OK". This will save the information in plaintext format.
11. Exit from your browser, open your spreadsheet, and import the file.
12. To work with the data, you will have to convert the plaintext dates to date format and numerical values to numbers. If you have EXCEL 5.0 or higher, the "Text Wizard" "General" format command does all the conversions appropriately.
13. Save the converted file in appropriate spreadsheet format (`*.xls for EXCEL`).

voir regulation on FDCs is discussed in Section 10.2.5.

Significance

The streamflow rate is itself highly significant as an indicator of water availability for instream and withdrawal uses, and many other quantities (e.g., water-quality determinants such as dissolved-oxygen, dissolved-solids, and suspended-sediment concentrations; stream-habitat determinants such as velocity and depth; and stream erosive power) are at least partly dependent on flow rate. Thus FDCs are extremely versatile and useful tools for analysis of water-resource problems, as described in detail in Section 10.2.5.

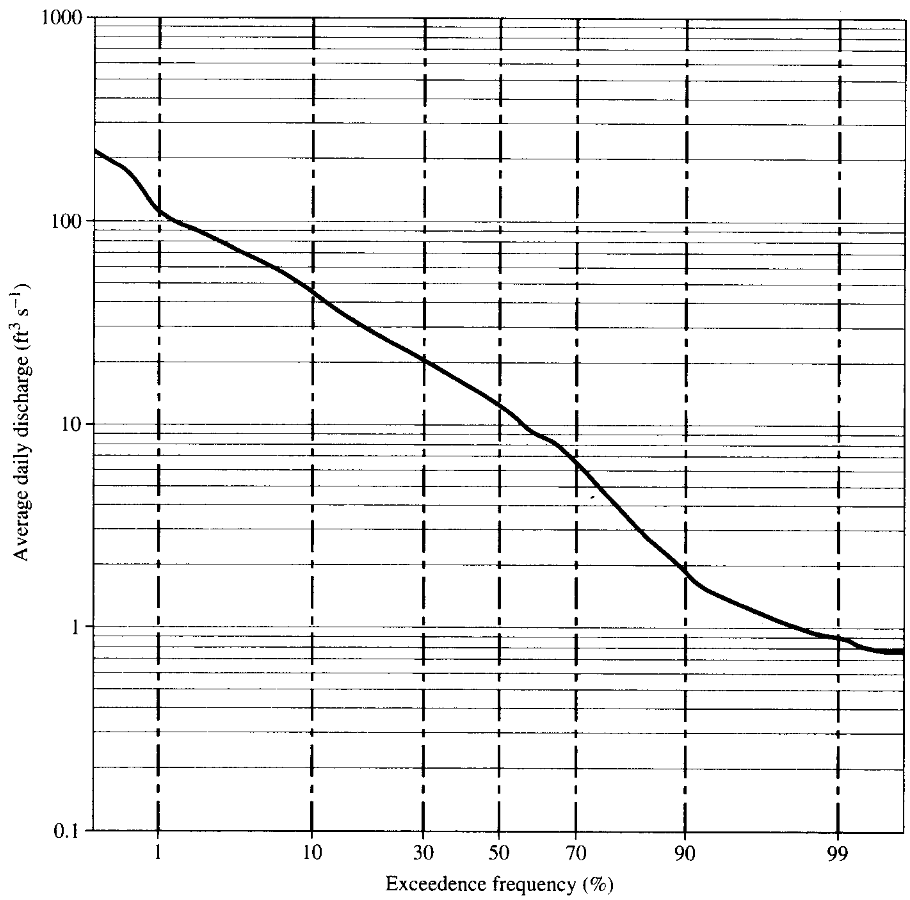
2.8 STORAGE, STORAGE EFFECTS, AND RESIDENCE TIME

2.8.1 Storage

The control volumes in Figures 1-1 and 1-2 represent storage in the hydrologic cycle, and the entire watershed or region within the dashed boundaries can also be thought of as a storage reservoir. The term "storage" often connotes a static situation, but, in reality, water is always moving through each control volume. In fact, it can be said that *water in the hydrologic cycle is always in motion AND always in storage.*

In many hydrologic reservoirs, such as lakes, segments of rivers, ground-water bodies, and water-

FIGURE 2-7
Flow-duration curve of mean daily flows for the Oyster River near Durham, NH, plotted on logarithmic-probability paper.



sheds, the outflow rate increases as the amount of storage increases.[8] For these situations, we can model the relation between outflow rate, q, and storage volume, S, as

$$q = f(S). \tag{2–25}$$

In some cases, the nature of the function $f(S)$ can be developed from the basic physics of the situation; in others, such as natural watersheds, Equation (2-25) is merely a conceptual model. The simplest version of Equation (2-25) describes a **linear reservoir**:

$$q = k \cdot S, \tag{2–26}$$

where k is a positive constant. Although no natural reservoir is strictly linear, Equation (2-26) is often a useful approximation of hydrologic reservoirs. It is used as the basis of the "convex watershed model" in Section 9.1.5.

2.8.2 Storage Effects

Where Equation (2-25) applies, storage has two effects on outflow time series:

1. It decreases the **relative variability** of the outflows relative to the inflows. Standard statistical measures such as the coefficient of variation (ratio of standard deviation to mean; Section C.2.5) or simple ratios determined from FDCs (e.g., $q_{.50}/q_{.95}$) can be used to characterize relative variability quantitatively.

2. It increases the **persistence**—the tendency for high values of a time-distributed variable to be followed by high values, and low values by low values—of the outflow time series relative to the inflow time series. As is explained in Section C.9, persistence can be characterized by the **autocorrelation coefficient** of a time series.

[8]Note that there are many hydrologic reservoirs for which this relation does *not* hold, e.g., a melting snowpack, the global atmosphere, the global ocean.

2.8.3 Residence Time

Residence time, or **average transit time**, is a universal relative measure of the storage effect of a reservoir. It is equal to the average length of time that a "parcel" of water spends in the reservoir. If outflow rate and storage are related as in Equation (2-25), the relative effect of the storage on the variability and persistence of the outflows increases as residence time increases. Figure 2-8 illustrates these effects for linear reservoirs [Equation (2-26)].

Residence time can be calculated by dividing the average mass (or volume) of the substance of interest in the reservoir, S, by the average rate of outflow, m_Q, or inflow, m_I, of that substance:

$$T_R = \frac{S}{m_Q} = \frac{S}{m_I}, \tag{2–27}$$

where T_R is residence time. Note that Equation (2-27) holds only when $m_Q = m_I$ —that is, when the average rate of change of storage is 0; as noted above, this assumption can usually be made for natural reservoirs over the long term. Note also that Equation (2-27) is dimensionally correct. Residence time is also called **turn-over time**, because it is a measure of the time it takes to completely replace the substance in the reservoir.

For many hydrologic reservoirs, such as lakes, values of S and of m_Q or m_I can be determined readily, and computation of residence time is

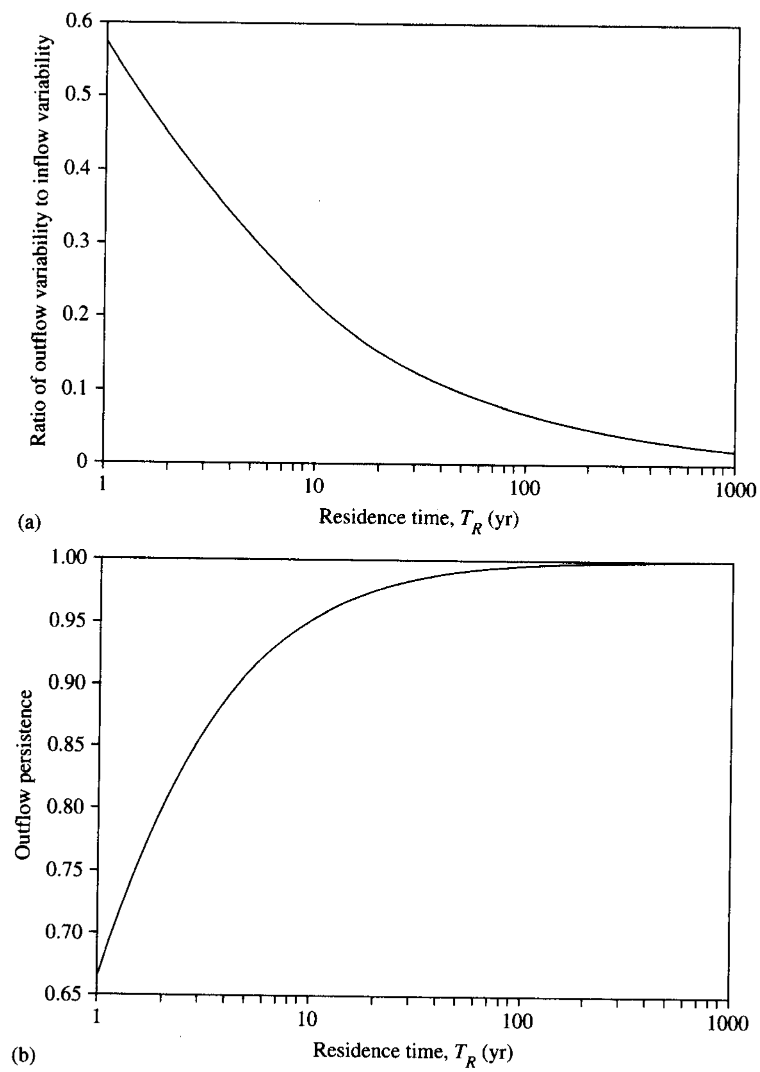


FIGURE 2-8
(a) Ratio of relative variability of outflows to relative variability of inflows as a function of residence time for a linear reservoir [Equation (2-26)]; and (b) Persistence of outflows (expressed as the autocorrelation coefficient) as a function of residence time for a linear reservoir when inflows have no persistence.

straightforward. For others, such as watersheds and ground-water bodies, it may be difficult to determine the value of S with precision; in these cases, we can usually speak of residence times in relative terms—for example, under natural conditions, outputs from most ground-water reservoirs enter streams and provide the bulk of the streamflow, at least between rain and snowmelt events (Sections 8.5.3 and 9.1). Thus, under similar climatic regimes, streams receiving water from ground-water reservoirs with large residence times (i.e., with large volumes of storage per unit watershed area) will tend to have less variable and more persistent streamflow than those receiving water from reservoirs with shorter residence times.

2.9 HYDROLOGIC MODELING

Much current research in hydrology is directed at improving our ability to predict or forecast the effects of land-use and climate changes on the water balance, ground-water levels, streamflow variability, and water quality of regions ranging from hillslopes or landfills to river basins to entire continents. These, and most other applications of hydrology to practical problems of design and forecasting, require the use of hydrologic models. A principal motivation for understanding the physics of hydrologic processes as developed in this text is to provide a sound basis for development and application of such models.

This text explicitly discusses the modeling of snowmelt (Section 5.5), infiltration (Section 6.6), interception (Section 7.6.3), evapotranspiration (Section 7.8), ground-water flow (Section 8.1.6), open-channel flow (Section 9.3), and runoff to streams (Section 9.5). Although our focus will be on the science underlying such modeling, the centrality of modeling to the study and application of hydrology requires that we explore approaches to and issues in modeling more broadly, and that exploration is the goal of this section.

We begin with a consideration of what a hydrologic model is; then we consider how and why models are used in hydrologic science and engineering, the various types of models, and the modeling process. The discussion concludes with an introduction to the BROOK90 model, which links the land-surface hydrologic processes that are the focus of this text into an integrated watershed model that can be used for exploring the impacts of climate and land use on regional hydrology.

2.9.1 What Is a Model?

In this text, the term "model" will refer to **simulation models**. Such a model is a representation of a portion of the natural or human-constructed world "which is simpler than the prototype system and which can reproduce some but not all of the characteristics thereof" (Dooge 1986). The essential feature of a simulation model is that it produces an output or a series of outputs in response to an input or series of inputs. The characteristics and use of the three major classes of simulation models that have contributed to scientific and applied hydrology are described in the following paragraphs.

A **physical model** is a tangible constructed representation of a portion of the natural world. If it is constructed at a larger or smaller scale than the natural system, formal rules of scaling based on dimensional analysis (see Appendix A; King et al. 1960) are used to relate observations on the model to the real world. Physical models have been important means to understanding problems of hydraulics and fluid mechanics, and they are often used to help design complex engineering structures, particularly those involving open-channel flow. Ground-water hydrologists have used physical models to simulate two-dimensional ground-water flow under various boundary conditions. One-to-one-scale physical models in the form of sprinkler-plot studies have been used to understand the process of infiltration (e.g., Nassif and Wilson 1975; see Figure 6-19), and small-scale physical models of watersheds have been used to elucidate some basic characteristics of watershed response to rainfall (Amorocho and Hart 1965; Grace and Eagleson 1966; Dickinson et al. 1967; Chery 1967, 1968; Black 1970, 1975).

Analog models use observations of one process to simulate a physically analogous natural process. For example, the flow of electricity as given by Ohm's Law is exactly analogous to Darcy's Law of ground-water flow, so the distribution of electrical potentials (voltage) on specially designed conductive paper can be used to determine the patterns of ground-water potentials (and hence of ground-water flow) under various boundary conditions.

A **mathematical model** is an explicit sequential set of equations and numerical and logical steps (or rules or "recipes") that converts numerical inputs representing flow rates or states of storages to numerical outputs representing other flow rates or storage states. The "guts" of a mathematical model are the equations whose forms represent the qualitative behavior of the flows and storages and the **parameters**—numerical constants—in these equations that dictate the quantitative behavior.

As the availability of more powerful digital computers, modeling techniques, and software has rapidly increased, the use of both physical and analog models in hydrology has been largely replaced by that of computer-implemented mathematical models, which are usually cheaper and much more flexible. Such models are usually mathematical representations of 'box-and-arrow' diagrams similar to those shown in Figures 1-1 and 1-2. Unless otherwise stated, our subsequent discussion of models will focus on mathematical models.

Perhaps the best metaphor for hydrologic models of all types is a map: A model is to hydrologic reality as a map is to the actual landscape. A mental comparison of a map of a region you're familiar with to the actual region gives a good sense of how a model approximates reality. The map metaphor also makes clear two essential characteristics of models:

1. *A model, like a map, is designed for a specific purpose.* A model emphasizes features appropriate to its purpose, while omitting other features: a road map shows road types, route numbers, and locations of cities and towns, but usually does not show topography, land cover, or other features that might be extremely important for purposes other than finding your way by car from point A to point B. (Actually, topography might be very important for such a purpose, yet still be omitted from your map!)

2. *A model, like a map, is constructed at a particular scale.* Neither represents features that are not visible at that scale—and some of the omitted features could be important in many contexts.

The issue of scale in modeling is an active area of research and discussion in the hydrological literature (e.g. Giorgi and Avissar 1997; Bergstrom and Graham 1998).

2.9.2 Purposes of Models

Hydrologic simulation models are developed either (1) to guide the formulation of water-resource-management strategies (including the design of structures) or (2) as tools of scientific inquiry.

Virtually all applications of hydrology to practical water-resource problems involve the use of models. These applications require either predictions or forecasts:

Predictions are estimates of the magnitude of some hydrologic quantity (e.g., the peak flow) that is either (1) associated with a particular exceedence probability or statistic of the quantity or (2) produced by a hypothetical rainfall or snowmelt event (often called the **design storm**).[9] Predictions are the basis for the design of civil engineering works such as reservoirs and reservoir spillways and of land-use plans (e.g., floodplain zoning) and for the assessment of the hydrologic impacts of land-use and climate changes.

Forecasts are estimates of the response to an actual anticipated event; e.g., the peak flow rate that will result from the rain that is expected in the next 24 hr on a given watershed. Forecasts are used to guide the operation of reservoir systems and to provide flood warnings to floodplain occupants.[10]

In the scientific context, models are used along with data to test hypotheses about the processes operating in some portion of the hydrologic cycle (Beven 1989). Models are developed to give explicit form to concepts of hydrologic function, and comparison of modeled hydrologic response with observed responses might confirm, or suggest revision

[9]Design storms are often specified as the event with a particular exceedence frequency (i.e., by a quantile of the input event such as the 10-yr, 1-hr rainfall on a given watershed). Approaches to estimating exceedence probabilities of rainfalls of specified durations are discussed in Section 4.4.4.

[10]Forecasting models are also commonly used for **hindcasts** (or **backcasts**), in which the objective is to estimate an unmeasured hydrologic response to a past event.

of hypotheses about the importance of various processes or the ways in which those processes are related. For example, you might assume, from a priori knowledge of the topography and geology of a watershed, that ground-water outflow (deep seepage) is not significant, develop a model of runoff processes that omitted deep seepage, and find that measured streamflow was consistently less than that predicted by the model. This might lead you to conduct further field studies to identify the reason for the discrepancy and perhaps ultimately to revise your concept of watershed processes by including ground-water outflow. Wigmosta and Burges (1997) reported an excellent example of the interactive use of models and field measurement in understanding runoff processes.

Development of hydrologic models has been a natural consequence of the widespread need for predictions and forecasts, along with (1) the complexity and spatial and temporal variability of hydrologic processes and (2) the limited availability of spatially and temporally distributed hydrologic, climatologic, geologic, pedologic, and land-use data.

2.9.3 Types of Models

Any given model can usually be described by one or more terms from each of the six categories in Table 2-3. The terms denoting physical domain and process should be reasonably self-explanatory (and will become more so as you progress through this text); those in the other categories are defined in Box 2-3.

2.9.4 The Modeling Process

The modeling process is diagrammed in Figure 2-9; its major elements are (1) conceptualization of the problem, (2) selection or development of the appropriate model, (3) parameter estimation ("calibration"), and (4) acceptance testing ("validation").[11]

[11] Matalas and Maddock (1976) suggested that the terms "parameter estimation" and "acceptance testing" are preferable to "calibration" and "validation," respectively, because they "remind us that a model is an abstraction of the physical process and not the physical process per se".

Conceptualization of the Problem

The most important element in the modeling process is the determination of the overall form and essential components of the model. These decisions must be based on a clear idea of the scientific or engineering purpose of the model, and this idea must be translated into an explicit formulation of the nature and form of the model output that is required—specifically,

- the type of information required (e.g., peak flows, flow volumes, ground-water heads, soil-water contents, evapotranspiration rates);
- the required accuracy and precision of the output;
- the locations for which the output is required;
- the time intervals for which the output is required.

The model conceptualization is dictated also by the nature and form of the information that is available about the system being modeled and the nature and form of the available input data—and, we must not forget, by the resources and time available to collect needed information that is not already available.

Model Selection or Development

The hydrologic literature abounds with descriptions of models; once the basic model requirements are established, one can usually develop the requisite software by implementing approaches developed by others with modifications to apply to the situation of interest. Many models are readily available as computer software designed to be easily modified to apply to a particular situation; descriptions of many of these were given by DeVries and Hromadka (1992) for streamflow and watershed models and by Anderson et al. (1992) for subsurface-flow models.

Parameter Estimation and Acceptance Testing

Both parameter estimation and acceptance testing require measured values of input and output quantities for the prototype system of interest and involve numerical and/or graphical comparison of measured outputs to modeled outputs. This requires splitting the measured data into a "parame-

TABLE 2-3
Terms Used to Characterize Hydrologic Models. See Box 2-3 for definitions of terms.

Physical Domain	Process	Simulation Basis
Vegetative canopy	Interception	Physically based
Snowpack	Snowmelt	Conceptual
Unsaturated zone	Infiltration	Empirical/regression
Aquifer	Overland flow	Stochastic—time series
Hillslope	Unsaturated flow	
Stream reach	Transpiration	
Stream network	Ground-water flow/head	
Lake or reservoir	Evaporation	
Watershed	Open-channel flow	
Region/continent	Stream hydrograph	
	Integrated watershed/region	
Spatial Representation	**Temporal Representation**	**Method of Solution**
Lumped	Steady state	0-dimensional
Distributed	Steady state—seasonal	Formal-analytical
Coordinate system	Single event	Formal-numerical
1-dimensional	Continuous	Finite difference
2-dimensional		Finite element
3-dimensional		Other
		Hybrid

ter-estimation set" and an "acceptance-testing set." There are no firm rules for the proportion of the total data allocated to each set, but usually no more than half the data is allocated for acceptance testing. For some types of data the allocation can be done randomly, but for data representing a time sequence it is usually necessary to select a continuous period for parameter estimation and a prior or subsequent period for acceptance testing.

In both parameter estimation and acceptance testing, it is important to focus on the purposes of the model. For example, the following aspects of model output might be more or less important in different contexts:

- the ability to reproduce the long-term or spatial mean value of a quantity;
- the ability to reproduce overall variability (e.g., the standard deviation or range of the quantity);
- the ability to minimize *actual errors* (errors as measured in the units in which the quantity is measured; e.g., streamflows in m^3s^{-1});
- the ability to minimize *relative errors* (errors expressed as a percentage of the mean);
- the ability to reproduce high values of a quantity (e.g., peak streamflows);
- the ability to reproduce low values of a quantity (e.g., drought streamflows);
- the ability to reproduce patterns of seasonal or spatial variability.

It is also important to remember that measured values of model inputs and outputs are themselves more or less subject to errors due to instrumental limitations and the inherent inability of observational networks like rain gages or wells to capture the spatial variability of input or output quantities. To the extent that such errors exist, parameter selection and evaluation of model performance will be subject to error.

Parameter Estimation The objective of parameter estimation is to determine appropriate values for model parameters whose values are not known a priori—for example, hydraulic conductivity in Darcy's Law of ground-water flow [Equation (8-1)] or the proportionality constant in the linear-reservoir equation [Equation (2-26)]. The input data of the parameter-estimation set are entered into the model and the values of parameters are systematically adjusted to determine which values give the "best" fit between the modeled and the measured outputs according to predetermined criteria. Fit can be judged qualitatively by visual comparison of measured and simulated hydrographs or flow-duration curves or of scatter plots of simulated vs. actual output quantities. Numerical measures of best fit are reviewed by Martinec and Rango (1989) and discussed in Section C.10.

BOX 2-3

Definitions of Modeling Terms in Table 2-3

Simulation Basis

Physically Based Uses equations derived from basic physics [e.g., conservation of mass, energy, or momentum; force balance; diffusion (see Table 2-1)] to simulate flows and storages.

Conceptual Uses "reasonable" a priori relationships to simulate flows and storages. Example: Outflow from storage modeled as proportional to the amount of water in storage [Equation (2-26)].

Empirical/Regression Uses approximate relationships developed from observations to simulate flows and storages. Statistical regression techniques are often used to develop these relationships. Example: Snowmelt rates modeled as proportional to air temperature [Equation (5-57)].

Stochastic Time Series Uses formal time-series-analysis techniques to characterize the behavior of flows and storages. Example: Estimation of ground-water recharge via impulse-response analysis [Equation (8-39); see Salas (1992) for a review of these techniques.]

Spatial Representation

Lumped Region or watershed represented as a point. Spatially varying inputs, soils, vegetation, topography, and so on are each characterized by a single parameter that is an average or representative value. Also called '0-dimensional' models.

Distributed Provides some representation of the spatial variability in a region. Can range from dividing the region into two subregions (e.g., upland and lowland) to more elaborate representations of spatial variability (e.g., subdividing the region by a grid system, with model parameters that vary from cell to cell). Sometimes accomplished by linking lumped models.

Coordinate System Formal coordinate system of 1, 2, or 3 dimensions is used to represent space. Used where formal mathematical relations are basis for model. Coordinate system is usually orthogonal (Cartesian), but radial coordinates are used for ground-water models involving flows to or from wells.

Temporal Representation

Steady State Outputs are one or more values that represent a long-term average or the ultimate magnitude of a quantity. Example: Global average temperature, simulated as response to changes in solar output or changes in land or cloud cover (Box 3-1).

Steady-State Seasonal Outputs are long-term seasonal averages of a quantity. Example: The Thornthwaite water-balance model (Box 7-3) predicts average monthly values of evapotranspiration in response to average monthly values of precipitation and temperature.

Single-Event Simulates time-varying response of a system to an isolated input. Example: SCS model (Section 9.6.2) simulates streamflow response to a single design storm.

Continuous Outputs are a sequence of responses to a sequence of inputs over a specific period. The time step of the sequence may be days, months, years, or other periods. Example: The BROOK90 model (Section 2.9.5) simulates one or more years of daily hydrologic responses to specified daily weather data.

Method of Solution

0-Dimensional Computations not based on formal coordinate system, usually employed in lumped models. Examples: Thornthwaite model of evapotranspiration (Box 7-3); SCS flood-hydrograph model (Section 9.6.2).

Formal–Analytical Basic differential equations in coordinate system solved *analytically*. Example: Philip solution of Richards Equation of infiltration [Equation (6-31)].

Formal–Numerical Basic differential equations in coordinate system solved by *finite-difference* or *finite-element* discretization schemes. (See Wang and Anderson 1982; Bear and Verruijt 1987.) Example: Regional ground-water flow models (Section 8.2).

Hybrid 0-dimensional and formal solutions used for different processes within model. Example: BROOK90 model (Section 2.9.5) uses formal–numerical solutions for soil-water movement (Box 6-3) and 0-dimensional methods for other processes (Box 4-1, 5-1, 7-2, 7-4).

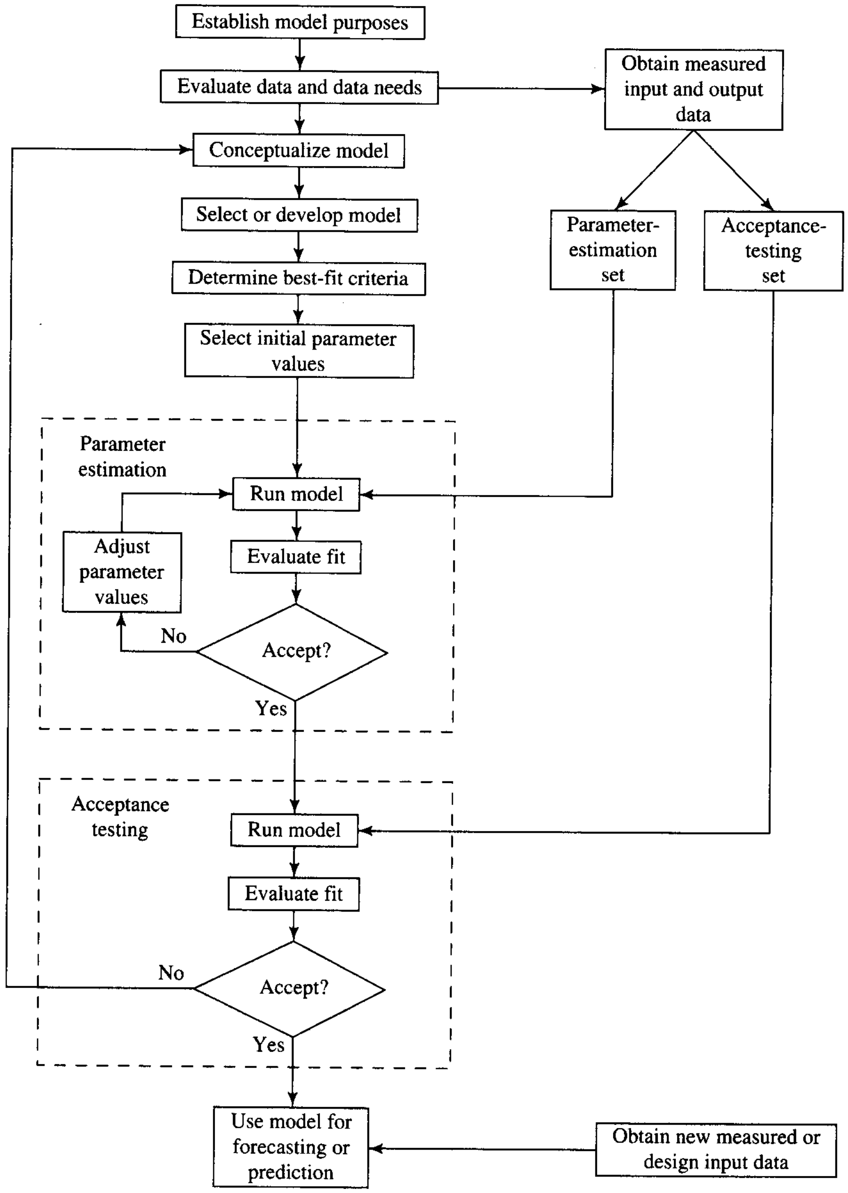


FIGURE 2-9
Flow chart for the modeling process.

Although conceptually straightforward, the parameter-estimation process is often fraught with difficulty and ambiguity, especially in multiparameter models:

- Very different sets of parameter values may give nearly equivalent fits.
- Model outputs may be insensitive to the values of one or more parameters.
- One or more best-fit values may differ greatly from what seems intuitively reasonable.
- Best-fit values may differ in different time periods.

When these situations occur, confidence in a model's ability to simulate the situation of interest is diminished.

Acceptance Testing Once the parameters are selected, performance testing leading to acceptance or rejection of the model for a particular application should be evaluated by graphical and/or numerical comparison of modeled and measured outputs *for situations not used in parameter estimation.* As an example, Figures 2-10 and 2-11 compare streamflows simulated by the BROOK90 model (described in Section 2.9.5) with measured values

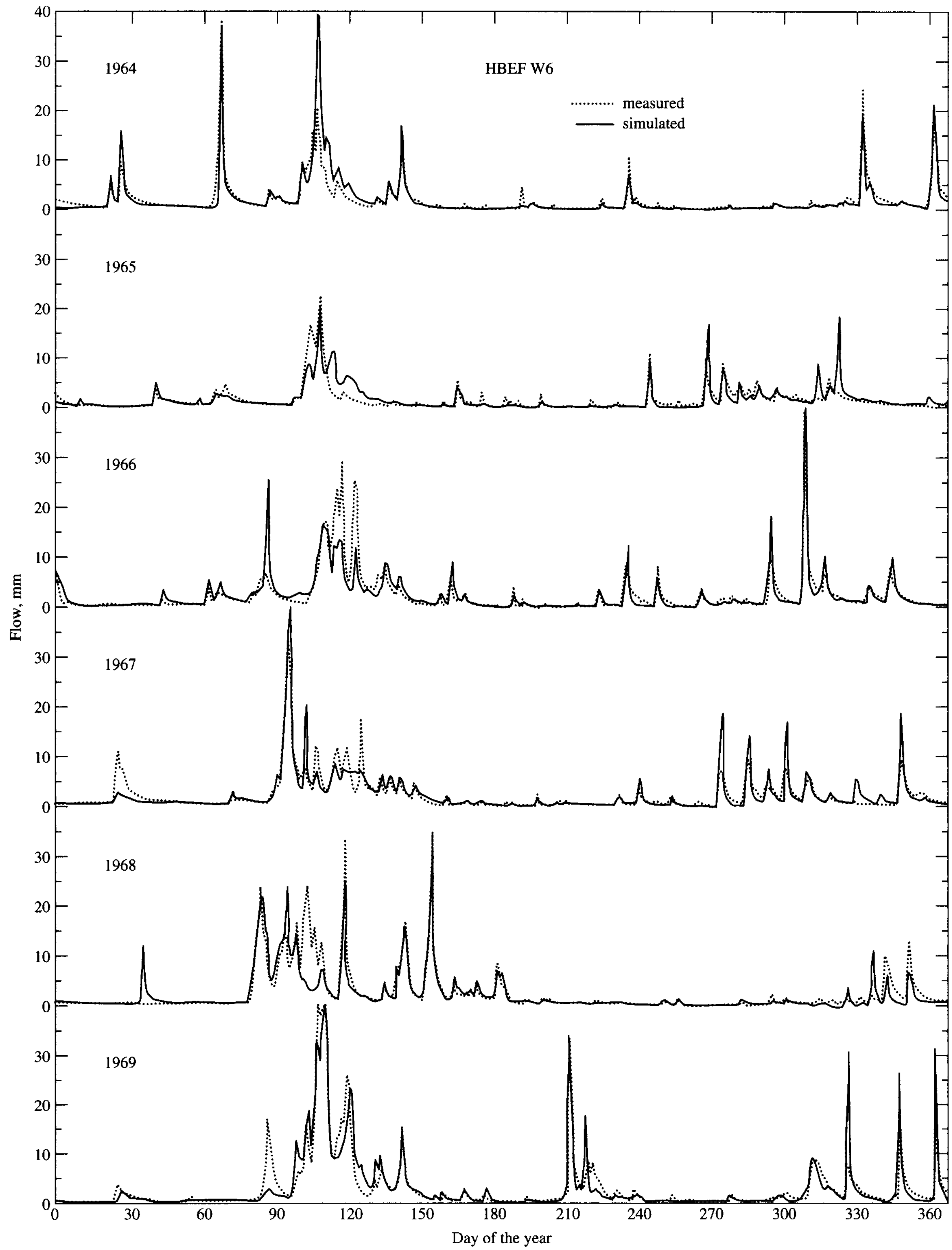


FIGURE 2-10
Comparison of measured daily flows at Hubbard Brook Experimental Forest, West Thornton, NH, (dotted lines) with flows simulated by BROOK90 (solid lines) for 6 years. From Federer (1995).

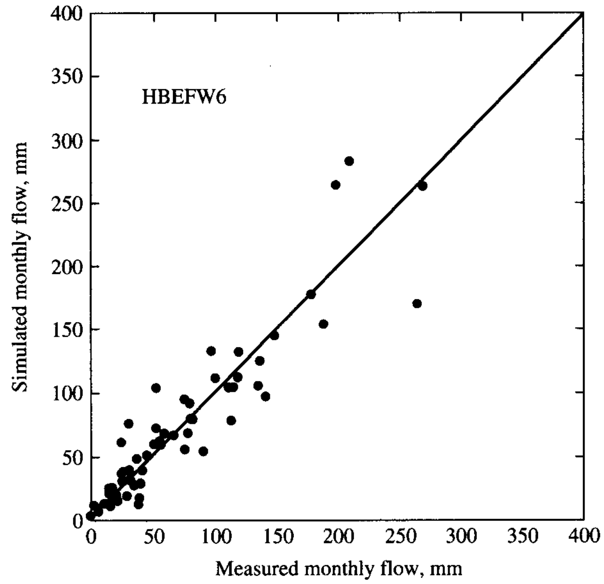


FIGURE 2-11
Scatter plot of monthly flows simulated by BROOK90 vs. measured flows for 1964-1969 at Hubbard Brook Experimental Forest, West Thornton, NH. Line shows 1:1 ratio. From Federer (1995).

in a research watershed. If a model does not satisfactorily simulate the measured values, a new model, perhaps based on a revised conceptualization of the situation, should be developed.

Klemeš (1986b) provided an excellent discussion of the philosophy and process of testing simulation models, and studies comparing models are published by the World Meteorological Organization (1986b, 1992) and Perrin (2001).

2.9.5 The BROOK90 Model

This book emphasizes the scientific understanding of individual processes (precipitation, snowmelt, infiltration, evapotranspiration, and others) in the land phase of the hydrologic cycle. In order to provide a means of synthesizing this understanding into a form that can be used to answer hydrologic questions, the text also describes a model called BROOK90. This model incorporates our scientific understanding into an integrated model that connects these processes to simulate the behavior of a watershed. We also give information about how BROOK90 can be accessed on the Internet (in Exercise 2-7) and how it can be used to predict watershed responses to climatic inputs and land-use changes. Here we introduce the basic features of BROOK90; a more detailed discussion of how the model simulates each specific process is given in the chapter that treats that process.

As can be seen in Table 2-3, BROOK90 is an integrated model that provides detailed simulations of the processes of interception, snowmelt, infiltration, unsaturated soil-water flow, transpiration, and soil evaporation and more approximately simulates overland flow and ground-water flow. Its structure is shown in Figure 2-12—note the similarity to Figure 1-2. It is largely a physically-based, spatially-lumped, continuous-time (daily time step) model that uses 0-dimensional computational approaches for processes other than soil-water movement; the latter is simulated by using a formal finite-difference numerical approach.

BROOK90 was developed by Dr. C.A. Federer, formerly of the U.S. Forest Service, who actively updates the model and maintains the BROOK Web site. The following overview is taken from Federer (1995):

> The hydrologic model BROOK90 simulates the water budget on a unit land area at a daily time step. Given precipitation at daily or shorter intervals and daily weather variables, the model estimates interception and transpiration from a single-layer plant canopy, soil and snow evaporation, snow accumulation and melt, soil-water movement through multiple soil layers, stormflow by source area or pipe-flow mechanisms, and delayed flow from soil drainage and a linear ground-water storage.
>
> BROOK90 simulates the land phase of the precipitation-evaporation-streamflow part of the hydro-

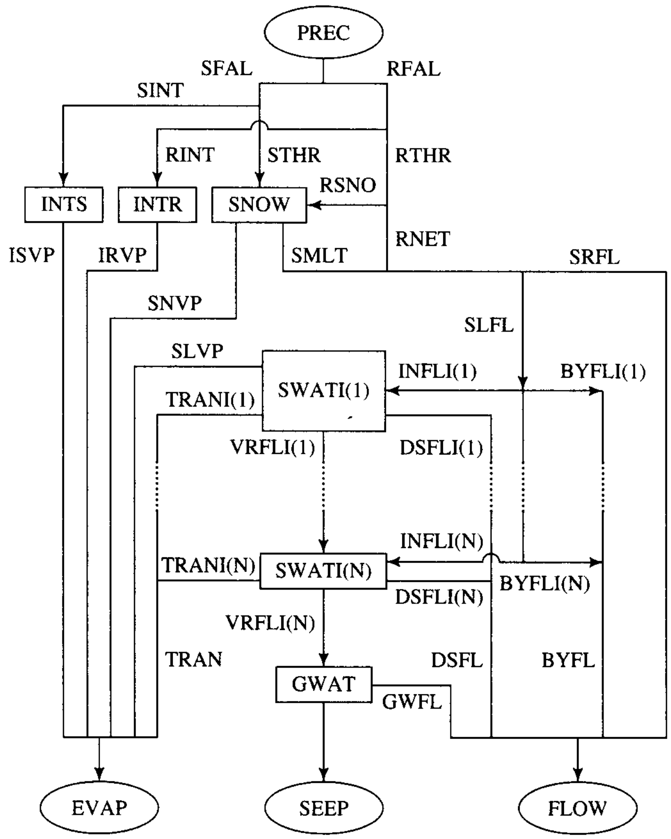


FIGURE 2-12
Flow chart for the BROOK90 model. Terms are defined below. Boxes are accumulated storages. All quantities expressed in mm of water. From Federer (1995).

PREC = Precipitation.
EVAP = Evapotranspiration.
SEEP = Ground-water outflow.
FLOW = Stream outflow.
INTS = Snow on vegetative canopy
INTR = Rain on vegetative canopy.
SNOW = Snowpack on ground.
SWATI = Soil-water storage (1 to N layers.)
GWAT = Ground-water storage.

logic cycle for a point or for a small, uniform (lumped parameter) watershed. There is no provision for spatial distribution of parameters in the horizontal. There is no provision for lateral transfer of water to adjacent downslope areas. Instead, BROOK90 concentrates on detailed simulation of evaporation processes, on vertical water flow, and of local generation of stormflow. Below ground, the model includes one to many soil layers, which may have differing physical properties.

BROOK90 has been designed to be applicable to any land surface. The model has numerous parameters, but all parameters are provided externally, are physically meaningful, and have default values. Parameter fitting is not necessary to obtain reasonable results. However, a procedure for modifying important parameters to improve the fit of simulated to measured streamflow is described.

BROOK90 is designed to fill a wide range of needs: as a research tool to study the water budget and water movement on small plots, as a teaching tool for evaporation and soil-water processes, as a water-budget model for land managers and for predicting climate-change effects, and as a fairly complex water-budget model against which simpler models can be tested.

Some published studies that have used current or earlier versions of the BROOK model include Federer and Lash (1978), Devillez and Laudelout (1986), Hornbeck et al. (1986), Hornbeck et al. (1987), Forster and Keller (1988), Yanai (1991), and Lawrence et al. (1995).

2.9.6 Final Words of Caution

Developing and working with computer simulation models is challenging and fun, and it can be done in a comfortable room with a coffee cup at hand. Collection of data in the field is also challenging, but often uncomfortable, tedious, frustrating, and expensive. Thus, although computer models have greatly facilitated the science of hydrology and its application and will play an increasing role in the future, we must continually remind ourselves that the goal is to understand

and predict nature, not to demonstrate our cleverness. As it was nicely phrased by Dooge (1986, p. 49S),

> Many . . . modelers seem to follow . . . the example of Pygmalion, the sculptor of Cyprus, who carved a statue so beautiful that he fell deeply in love with his own creation. It is to be feared that a number of hydrologists fall in love with the models they create. In hydrology, . . . the proliferation of models has not been matched by the development of criteria for the evaluation of their effectiveness in reproducing the relevant properties of the prototype.[12]

Models are essential tools for almost all practical applications of hydrology and can be powerful aids in scientific analysis. However, anyone seeking to use a model to provide predictions or forecasts that will be used for critical design or operational applications or scientific decisions should first review the discussions by Matalas and Maddock (1976), Klemeš (1986b), Dooge (1986), Beven (1993), Oreskes et al. (1994), and Perrin et al. (2001). The collective wisdom of these discussions can be summarized as follows:

> Although acceptable parameter values can be determined for almost any model, in most cases the parameters are not unique. In addition, because of the inevitable errors in measured data and the impossibility of representing the space-time continuum of nature as a finite array of space-time points, no model can be validated as a true simulation of nature.

EXERCISES

Exercises marked with ** have been programmed in EXCEL on the CD that accompanies this text. Exercises marked with * can advantageously be executed on a spreadsheet, but you will have to construct your own worksheets to do so.

***2-1** Using Equation (2-16) and assuming no model error, compute (a) the estimated evapotranspiration and (b) the absolute and relative uncertainties in the estimate for the following situations:

Location	a Connecticut River, USA	b Yukon River, Canada	c Euphrates River, Iraq	d Mekong River, Thailand
Watershed Area (km^2)*	20,370	932,400	261,100	663,000
Precipitation, m_P(mm yr^{-1})*	1,100	570	300	1,460
Relative Error in P, u_P^+	0.1	0.2	0.1	0.15
Streamflow, m_Q(m^{3}s^{-1})*	386	5,100	911	13,200
Relative Error in Q, u_Q^+	0.05	0.1	0.1	0.05

* Published values. $^+$Assumed - actual uncertainty unknown.

****2-2** Using the methods in Box C-1, compute the sample median and 0.25- and 0.75-quantile values of the three time series in Table 2-2. The data for this table are in file `Table 2-2.xls` on the CD accompanying this text.

****2-3** Using the methods in Box C-2, compute the mean, standard deviation, coefficient of variation, and skewness of the three time series in Table 2-2. The data for this table are in file `Table 2-2.xls` on the CD accompanying this text. Which time series is the most variable, relatively speaking? Which is the most skewed?

****2-4** Using the methods of Box C-5, compute the sample autocorrelation coefficients of the three time series in Table 2-2. The data for this table are in file `Table 2-2.xls` on the CD accompanying this text. Which time series shows the most persistence? (Hint: If there are N items in the time series, you can calculate the autocorrelation coefficient by using the spreadsheet's correlation function: specify the first data range as values 1 through $N - 1$, the second data range as values 2 through N.)

***2-5** Using the instructions in Box 2-2, obtain a time series of annual peak flows for a stream of interest. Using the methods of Box C-2, estimate the mean, standard deviation, and coefficient of variation of the time series.

2-6 (a) Obtain appropriate topographic maps and trace out the watershed of a stream of interest. (b) Measure the drainage area by using a digitizer, planimeter, or grid overlay. (c) What geologic information is available that could help you assess whether ground-water outflow is significant?

2-7 Access the "Compass Brook" home page on the World Wide Web at

[12]Note that quantitative criteria for evaluating models are discussed in Section C. 10.

http://users.rcn.com/compassbrook/compassb.htm

and click on "BROOK90". Read the information about the BROOK90 model and how to download it and obtain the documentation. Check with your instructor to see whether your institution has a site license and a copy of the documentation.[13] Read the first chapter of the documentation (Chapter B90). Follow the instructions to download Version 3.2*x* into a directory `B90V3_2` that you have created on your hard drive. (*x* designates the current version of the program; $x = 5$ at the time of writing.) In your File Manager, double click on `B90V3_2x.EXE` to execute the model. In the "Run time" box, input 2192 days. This will allow a continuous run for six years (1964–1969) for Watershed 6 (W6) at Hubbard Brook Experimental Forest (HBEF) in West Thornton, NH. Click on "Output" and select Annual, Monthly, and Daily values of *PREC, MESFL, FLOW, EVAP*, and *SNOW* (note all values are in mm). Click on "Run" "New run" and watch the hydrologic years 1964–1969 unfold. Identify what each color graph is showing and note the annual patterns of the quantities.

(a) Compare the curves representing simulated streamflow (*FLOW*) and measured streamflow (*MESFL*). Note the relative values of *FLOW* and *MESFL*. Are there consistent patterns in these values? What might cause the patterns observed?

(b) Make a scatterplot of monthly estimated streamflow (*FLOW*) vs. measured streamflow (*MESFL*). Assess the goodness of fit qualitatively by reference to the graphical printout. Supplement this qualitative analysis by computing the Nash–Sutcliffe criterion (Section C.10).

(c) Make separate scatterplots of *FLOW* vs. *MESFL* for each of the four seasons: Winter (Dec, Jan, Feb); Spring (Mar, Apr, May); Summer (Jun, Jul, Aug); Fall (Sep, Oct, Nov), and make the same kinds of qualitative and quantitative analyses as for the aggregated data. Which season gives the best results, and the worst? What do these results suggest about the model's shortcomings?

(d) Compute the water balance for each year (*PPTN, EVAP, MESFL*) and the change in storage at the end of each year. Note any differences between the balances for the wettest and driest years. Is the storage change at the end of each year significant?

(e) Qualitatively and quantitatively compare *FLOW* and *MESFL* for the wettest and driest years with the overall results of part (b).

(f) Compute *PPTN* − *MESFL* for all months and compare with monthly evapotranspiration (*EVAP*). Generate a time-series graph of both values and analyze the data for temporal patterns. Try to assess what accounts for these patterns.

(g) Describe and discuss the causes of the temporal patterns of soil moisture (*SWAT*) on the graphical printout.

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[13] You do not need a license for one-time use in an exercise, but *you should obtain a license if you plan to use the model regularly for educational or scientific purposes.* See the web site for instructions.