HYDROLOGY AND THE MANAGEMENT OF WATERSHEDS

A lake scene in northern Minnesota.

Fourth Edition

A waterfall in Guizhou Province, China, illustrates both the beauty and power of flowing water.

Fourth Edition

HYDROLOGY AND THE MANAGEMENT OF WATERSHEDS

Kenneth N. Brooks

Peter F. Ffolliott

Joseph A. Magner

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Dedicated to

John L. Thames, Hans M. Gregersen, and Leonard F. DeBano

and

Students and the People Who Manage Land and Water for Future Generations

CONTENTS

PREFACE

The fourth edition of this book is a major revision and restructuring of the earlier editions. The basic concepts and fundamental aspects of hydrology and hydrologic processes have been retained with the methods and applications of the science of hydrology in the management of watersheds. We have eliminated the separate chapters on snow hydrology, water resources development and engineering applications, and hydrologic methods contained in the earlier editions. The subject matter in these chapters has not been eliminated in this edition, but rather it has been integrated into chapters that emphasize the hydrologic processes, methods, and applications of integrated watershed management (IWM).

Given the accessibility of information through the Internet and the rapid advancement of technologies, we have referenced URLs throughout the book and presented them in **Webliographies** at the end of the chapters. This approach facilitates the acquisition of current methods, models, baseline data summaries, and applications of technologies for coping with today's challenging issues of changing land and water use and climatic variability.

The chapters in **Part 1 – Watersheds, Hydrologic Processes, and Pathways** – present an updated foundation for the study of hydrology and watershed management. Basic properties and principles of water and energy relationships on earth are presented that provide a basis for understanding the circulation, water form transformations, and flow processes that occur on watersheds. Subsequent chapters build on this basic foundation and concentrate on hydrologic processes and methods of measurement and analysis as presented in the earlier editions. Noted changes include topics of snow measurement and snowmelt processes covered in the chapter on precipitation; hydrologic methods of estimating streamflow characteristics are presented in the chapter on streamflow measurement and analysis; and an expanded chapter on groundwater that examines groundwater–surface water exchanges. The chapters in **Part 2 – Physical, Chemical, and Biological Linkages of Water Flows** – have been structured to reflect the current understandings of soil erosion processes and control of these processes; accumulation and movement of sediment in a stream channel; and the hydrologic linkages to water quality. A new chapter emphasizing geomorphology, valley and channel forming processes, evaluation, and classification is also included in this part of the book. **Part 3 – Integrated Watershed Management** – combines and updates much of the information presented in the third edition of the book into chapters on the management of forests, woodlands, rangeland watersheds for maintaining and where possible enhancing the flows of high-quality water from watersheds. The separate chapters on riparian systems and wetlands found in earlier editions of the book have been combined in a chapter on managing these ecosystems within the context of a watershed landscape. A chapter on the effects of fragmenting wildland watersheds into agricultural and urban areas; implementation of "best management practices"; the importance of regulatory compliance, and coping with climatic variability are contained in this part of the book. Socioeconomic considerations of watershed management are presented in a chapter. A chapter on tools and emerging technologies available to managers for more efficient and responsive watershed management concludes the book.

This fourth edition of the book is intended largely for introductory college courses in hydrology and watershed management. However, the book can also serve as a reference for personnel of governmental and nongovernmental organizations with responsibilities for the management of land, water, and other natural resources on watershed landscapes. The book is also suited for international audiences with examples of watershed processes and the management of land and water extending beyond the US borders. Examples of applications are liberally presented in the text and in boxes throughout the book to help students understand how basic principles and methods can be applied in practice.

Metric (SI) units are used in the book with the exception of where original formulas, figures, tables, and other unit-dependent relationships were developed originally in English (customary) units and where the conversion to metric units is awkward. A table of metric to English units is presented in an appendix to assist the reader in making conversions if desired.

The authors thank Clara M. Schreiber once again for her dedicated work in the preparation of this fourth edition. Clara has also been a partner in all of the earlier editions of the book. The contributions of Hans M. Gregersen, the late John L. Thames, and Leonard F. DeBano – collaborating authors of the earlier editions of this book – have been integral to the evolution of this fourth edition and are greatly appreciated. The contributions of Mark Davidson, Britta Suppes, Linse Lahti, Mary Presnail, Peter Magner, and Dain Brooks in providing new figures and photographs are appreciated. They have improved the visual presentations of the book.

DEFINITION OF TERMS

Hydrology is the science of water concerned with the origin, circulation, distribution, and properties of waters of the earth.

Forest Hydrology/Range Hydrology/Wildland Hydrology refer to branches of hydrology that deal with the effects of vegetation and land management on water quantity and quality, erosion, and sedimentation in the respective settings.

A **watershed or catchment** is a topographically delineated area drained by a stream system; that is, the total land area above some point on a stream or river that drains past that point. A watershed is a hydrologic unit often used as a physical-biological unit and a socioeconomic-political unit for the planning and management of watershed resources.

River basin is similarly defined but is a larger scale. For example, the Mississippi River Basin, the Amazon River Basin, and the Congo River Basin include all lands that drain through those rivers and their tributaries into the ocean.

Integrated watershed management is the process of organizing and guiding land, water, and other natural resource use on a watershed to provide desired goods and services to people without affecting adversely soil and water resources. Embedded in the concept of integrated watershed management is the recognition of the interrelationships among land use, soil, and water, and the linkages between uplands and downstream areas.

Watershed management practices are those changes in vegetative cover, land use, and other nonstructural and structural actions taken on a watershed to achieve watershed management objectives.

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A lake scene in northern Minnesota.

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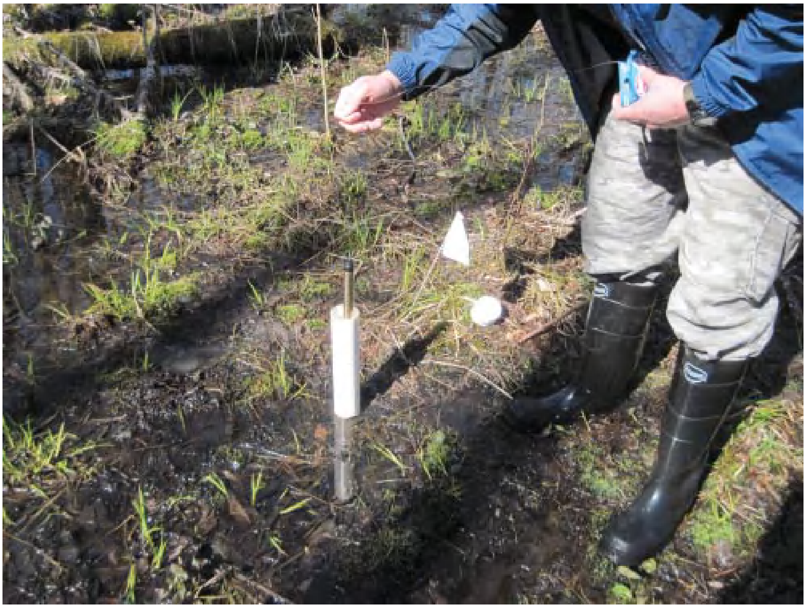

PLATE 1. Measuring groundwater levels in a forested wetland with a pressure transducer in a shallow well (Photograph by Chris Lenhart)

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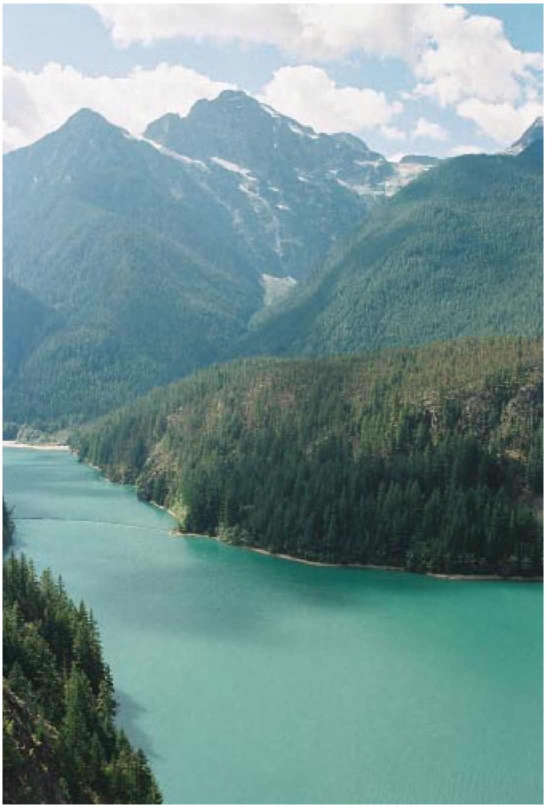

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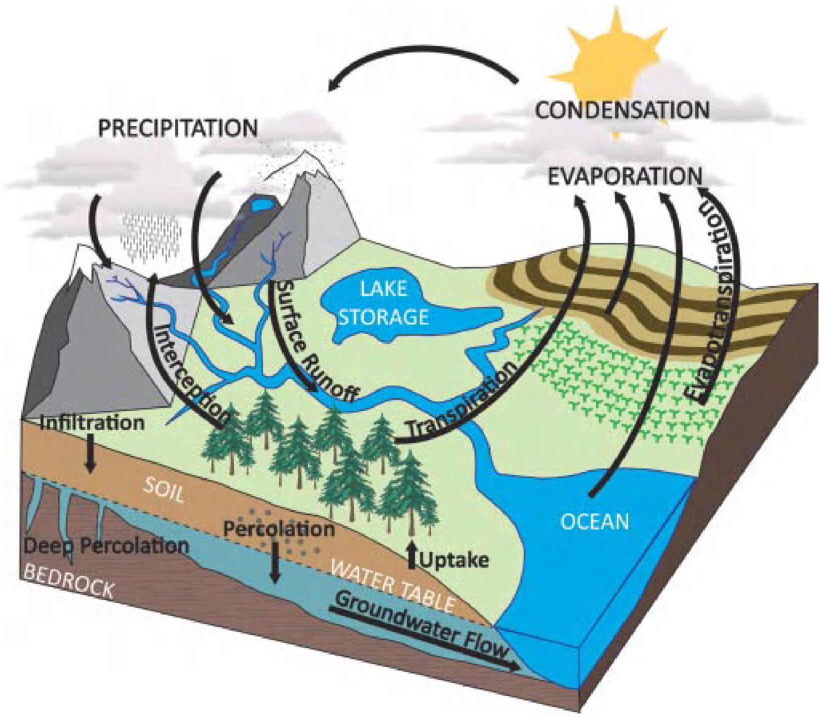
Plate 3. Students measuring streamflow with a current meter (Photograph by Lucas Bistodeau)



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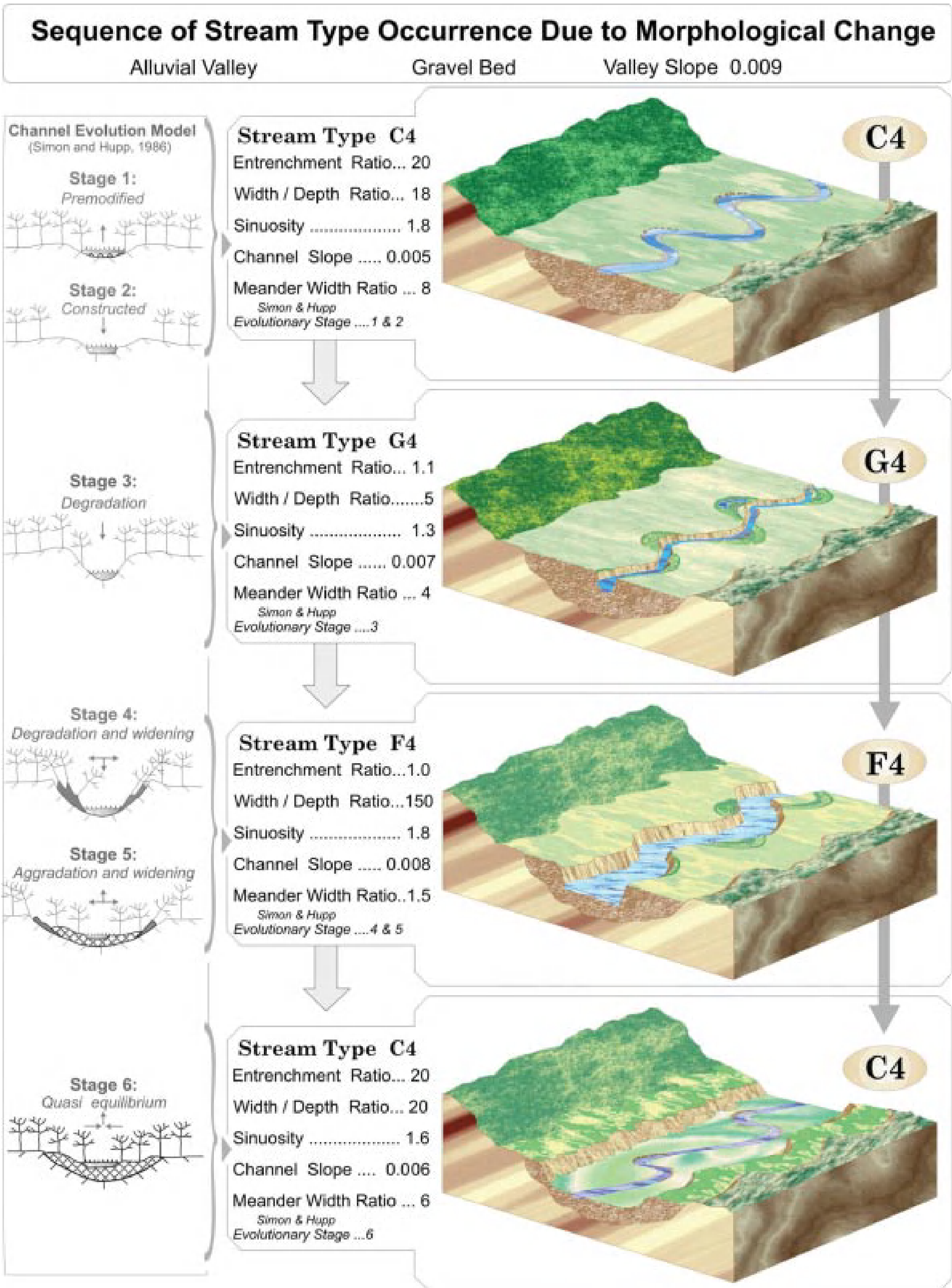

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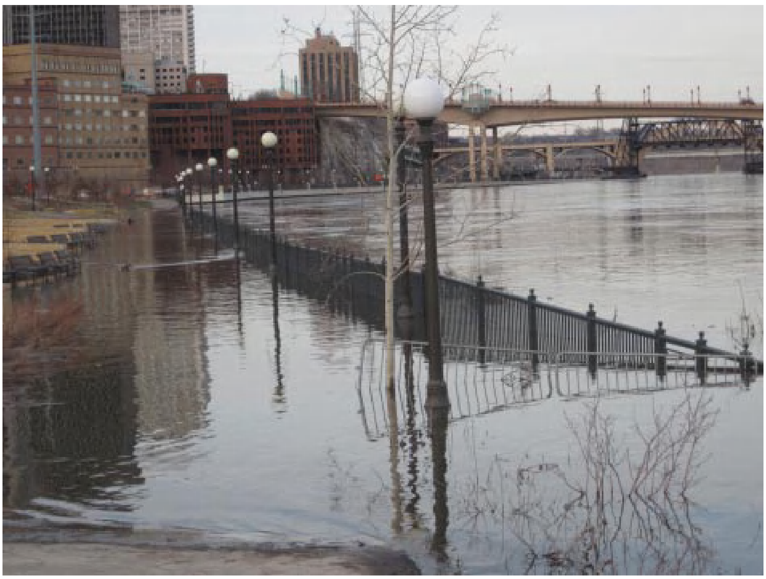

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PLATE 14. Undesirable effects on soils and water such as illustrated in this photo can be avoided when BMPs are properly implemented. In this case, a perennial vegetated buffer would provide some degree of protection for the water quality in this stream (Photograph by Mary Presnail)

PART 1

Watersheds, Hydrologic Processes, and Pathways

Photo 1. Measuring groundwater levels in a forested wetland with a pressure transducer in a shallow well (Photograph by Chris Lenhart) (For a color version of this photo, see the color plate section)

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Knowledge of the inherent characteristics of a watershed and hydrology provides the foundation for understanding the role of *integrated watershed management* (*IWM*) in achieving sustainable development and the use of land and water. An understanding of the hydrologic cycle and energy relationships on earth is fundamental to the study of hydrology and, therefore, necessary for making informed decisions in planning and implementing IWM practices. Embedded in this knowledge is an understanding of the nature of precipitation falling on the watershed; the magnitudes of evaporation, interception, and transpiration losses on a watershed; the infiltration and percolation of precipitation reaching the ground surface; and the pathways of water flow into stream channels and recharging groundwater aquifers. Methods of measuring or estimating streamflow discharges, including peak flows, minimum flows, volumes of flow, and the routing of streamflow from the watershed of origin to downstream points are also basic to understanding hydrology. Concepts of groundwater hydrology and processes of groundwater and surface water exchange are fundamental in understanding surface water–groundwater relationships in watersheds. Part 1 of this book focuses on obtaining this information.

Photo 2. Forested headwater watersheds are the source of most of the streamflow in the United States as depicted in this scene in the Northern Cascades of Washington (Photograph by Mark Davidson) (For a color version of this photo, see the color plate section)

Hydrologic processes are described in detail in Part 1. The topics presented in Chapter 1 of the book include a discussion on the inherent characteristics of a watershed; the importance of IWM; and fostering the sustainable use and development of natural resources while coping with land and water scarcity. Chapter 2 focuses on the hydrologic cycle, the water budget, the energy budget, and the energy processes that drive the hydrologic cycle. Precipitation, including rainfall and snowfall – primary inputs to the water budget, are considered in Chapter 3. The processes of evaporation, interception, and transpiration losses and their importance are discussed in Chapter 4. Infiltration processes and measurement and the pathways of water flow within a watershed system including groundwater recharge are the primary topics of Chapter 5. Methods of measuring and analyzing the streamflow response of watersheds are presented in Chapter 6. Basic concepts of groundwater and groundwater–surface water exchanges are presented in Chapter 7. The collective information in these chapters is basic to hydrology and provides the background necessary for a more comprehensive appreciation of how climatic factors, watershed characteristics, and land-use activities affect the hydrologic responses of watersheds.

PHOTO 3. Students measuring streamflow with a current meter (Photograph by Lucas Bistodeau) (For a color version of this photo, see the color plate section)

CHAPTER 1 *Introduction*

OVERVIEW

Our perspective of watershed management is that water and land resources must be managed in concert with one another. While hydrology and water quality are essential components of watershed management and are the subjects of much of this book, we also recognize the importance of ecosystem functions, land productivity, stream-channel morphology, and the actions of people on the land as integral parts of watershed management. To emphasize this holistic view, the concept of *Integrated Watershed Management* (*IWM*) is embedded in discussions presented in the chapters of this book. IWM deals not only with the protection of water resources but also with the capability and suitability of land and vegetative resources to be managed for the production of goods and services in a sustainable manner. Few watersheds in the world are managed solely for the production of water. Some municipal and power company watersheds that drain into reservoirs are the exception. Since water affects what we do on the landscape, watersheds serve as logical and practical units for analysis, planning, and management of multiple resources and coping with water issues regardless of the management emphasis.

A basic understanding of hydrology is fundamental to the planning and management of natural resources on a watershed for sustainable use. Hydrology enters explicitly and directly into the design of water resource projects including reservoirs, flood control structures, navigation, irrigation, and water quality control. Knowledge of hydrology also helps us in balancing the demands for water supplies, avoiding flood damages, and protecting the quality of streams, lakes, and other water bodies.

One of our concerns and an incentive for writing this book is that hydrology is not always considered in the management of forests, woodlands, rangelands, agricultural

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croplands, or in the array of human development activities on rural landscapes even though it should be! Ignoring development and land-use effects on soil and water resources is shortsighted and can lead to unwanted effects on a site and in downstream areas. For example, altering forested uplands, riparian communities, and wetland ecosystems can affect the flow and quality of water. Changes in vegetative cover that increase soil erosion can lead to soil instability and long-term losses of plant productivity. The consequences of soil erosion on upland watersheds can alter streamflow quantity and quality downstream. Changes in streamflow and sediment transport can, in turn, alter stream-channel morphology and affect the stability of rivers.

Hydrologic concepts and concerns about land use and water date back to some of the earliest recorded history. The evolution of hydrology from Egyptian texts as early as 2500 BC, to the ancient Indian writings from Vedic times before 1000 BC, to decrees recognizing the interrelationships between water and forests in Europe following the Dark Ages, to contemporary publications into the twenty-first century all illustrate the growing awareness of the importance of hydrology to the management of water and other natural resources. A more detailed timeline of the history of hydrology and watershed management is presented in Box 1.1.

Box 1.1

A Historical Look at Hydrology, Water, Watershed Management, and People

Year	US population[a]	Event	Reference
2125 BC	?	Canals move water from Nile River to supply water for irrigation and human consumption	Baines (2011) (www.bbc.co.uk)
1000 BC	?	An understanding of the hydrologic cycle was indicated in Indian texts from the Vedic times	Chandra (1990)
~900 BC	?	Chinese scholars develop an accurate understanding of the hydrologic cycle	Kittredge (1948)
~360 BC	?	Plato writes about land degradation, flooding, and erosion	Kittredge (1948)
1342	?	The first written record of a "protection forest" being established by a community in Switzerland	Kittredge (1948)

Year	US population[a]	Event	Reference
ca 1670	~110,000 (immigrants)	Perrault accurately quantifies the water balance of the Seine River watershed in France	Kittredge (1948)
1897	72 million	Organic Act . . . authorized the establishment of National Forests on public lands in the west	Glasser (2005)
1903	81 million	Gifford Pinchot publishes *A Primer of Forestry* in which he states "A forest, large or small, may render its service in many ways . . . especially against the dearth of water in streams."	Pinchot (1903)
1909	90 million	Wagon Wheel Gap paired watershed experiment initiated in Colorado	Bates and Henry (1928)
1912	95 million	Raphael Zon publishes *Forests and Water in Light of Scientific Investigation* as an appendix to the Report of the National Waterways Commission	Zon (1927)
1930s	123 million	Coweeta Hydrologic Laboratory founded near Asheville, North Carolina	Ice and Stednick (2004)
1950s and 1960s	151–179 million	Watershed studies established at H.J. Andrews, Oregon; Fool Creek, Colorado; Beaver Creek, Arizona; Fernow, West Virginia; Hubbard Brook, New Hampshire; Marcell, Minnesota; and other "barometer watersheds" on National Forests	Glasser (2005) Ice and Stednick (2004)
2010	308.7 million	US population; more than 1200 locally led watershed management districts, associations, partnerships, councils, and river basin commissions emerged for resolving watershed problems and achieving the goals of IWM	US Census Bureau

Source: US Census Bureau; 2010 population from website: http://2010.census.gov/2010census/data/
[a] Linearly interpolated from decadal census data.

Watersheds

Watersheds are biophysical systems that define the land surface that drains water and water-borne sediments, nutrients, and chemical constituents to a point in a stream channel or a river defined by topographic boundaries. Watersheds are the surface landscape systems that transform precipitation into water flows to streams and rivers, most of which reach the oceans. Watersheds are the systems used to study the hydrologic cycle (see Chapter 2), and they help us understand how human activities influence components of the hydrologic cycle.

Watersheds and Stream Orders

Watersheds and stream channels can be described according to their position in the landscape. It is useful to refer to an established nomenclature of stream orders (Horton, 1945; Strahler, 1964) in discussing watersheds and the water in streams that emanates from them. The commonly used method of stream orders classifies all unbranching stream channels as *first-order streams* (Fig. 1.1). A *second-order stream* is one with two or more first-order stream channels; a *third-order stream* is one with two or more second-order stream channels, and so forth. Any single lower stream juncture above a larger order stream does not change the order of the larger order stream. Thus, a third-order stream that has a juncture with a second-order stream remains a third-order stream below the juncture.

The watershed that feeds the stream system takes on the same order as the stream. That is, the watershed of a second-order stream is a second-order watershed and so on. While there is little evidence that streamflow and watershed characteristics are related to stream order, the use of this terminology helps one place a stream channel or a watershed in the context of the overall drainage network of a river basin. The physical and biological characteristics of watersheds and the climate in which they exist determine the magnitude

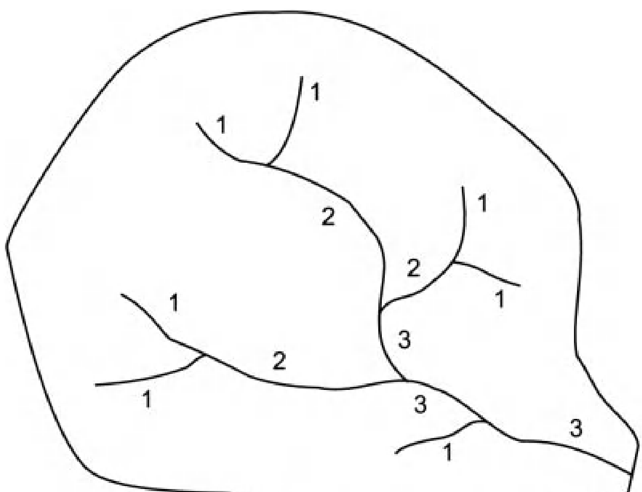

Figure 1.1. Stream order system by Horton (1945) as modified by Strahler (1964)

and pathways of water flow. Furthermore, the hierarchy of watersheds within a river basin generally influences the magnitude of water flow.

A Geomorphologic Perspective

As the upper-most watersheds in a river basin, first-order watersheds, also called *headwater watersheds*, are the most upstream watersheds that transform rainfall and snowmelt runoff into streamflow. Headwater streams comprise 70–80% of total watershed areas (Sidle et al., 2000) and contribute most of the water reaching the downstream areas in river basins (MacDonald and Coe, 2007). Headwater watersheds are often forested or once were prior to the expansion of agriculture, urban areas, and other human development activities. These headwaters are particularly important in water resource management. First-order streams in mountainous regions occur in steep terrain and flow swiftly through V-shaped valleys. High rainfall intensities can erode surface soils and generate large magnitude streamflow events with high velocities that can transport large volumes of sediment downstream. Over geologic time, mountains erode and sediment becomes deposited downstream (Fig. 1.2). As water and sediment from headwater streams merge with higher order streams, sediment is deposited over vast floodplains as rivers reach sea level. A transitional zone

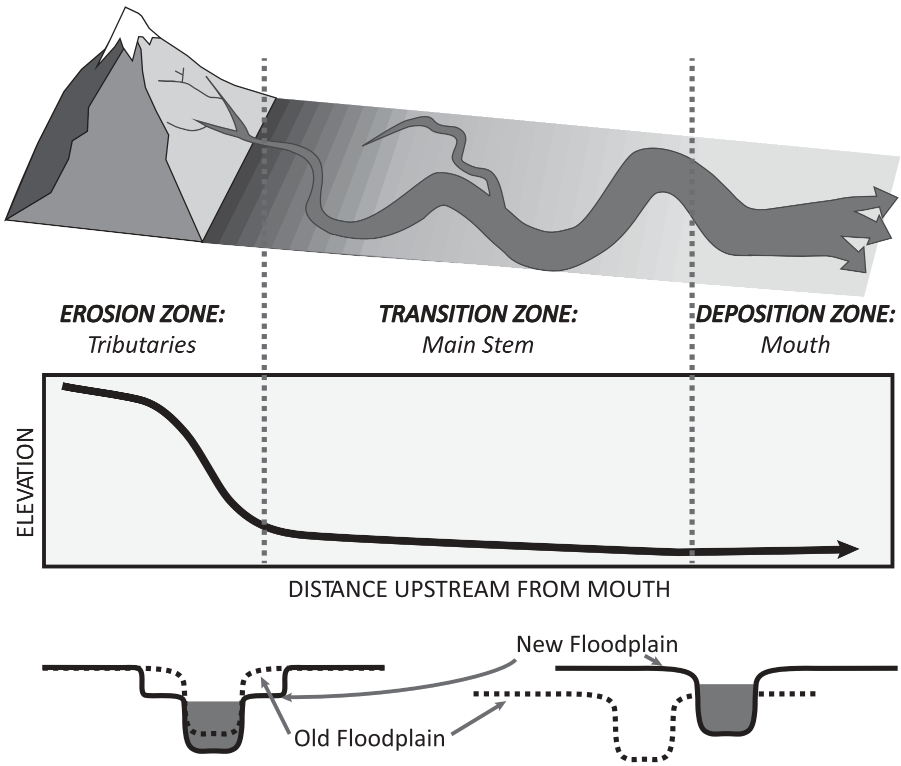

FIGURE 1.2. Rivers generally flow from an upper, high-gradient erosion zone through a transition zone to a low-gradient deposition zone (Schumm, 1977, as modified by Verry, 2007)

exists between the steep headwater steams and the lower zone of deposition at the mouth of major rivers and is typically characterized by broad valleys, gentle slopes, and meandering streams.

The "work" of water on soils, hillslopes, and within rivers forms landscapes with topography and soils that are better suited for some types of land use than others. Agricultural centers have developed in the transitional and depositional areas of a river basin while the steeper uplands are likely to prohibit intensive agricultural cultivation, resulting in landscapes with forests, woodlands, and rangelands suitable for forestry and livestock-grazing enterprises.

Watershed Assessments

The hydrologic response of watersheds to climatic variability and land-use changes will be discussed in this book requiring that methods of delineating watershed boundaries, determining watershed areas, and assessing a myriad of watershed metrics be understood. *Geographic Information Systems* (*GIS*), maps, and other tools, such as *Google Earth*, provide the means to quantify and describe watersheds, their vegetative cover, geology, and soils, and help delineate people's activities occurring across the landscape. Physical descriptions, such as watershed area, slope, stream-channel lengths, and *drainage density* (the sum of all channel lengths in a watershed divided by the watershed area), are important descriptors of watersheds.

Considerable information is available for assessing watersheds in the United States. *Hydrologic Unit Codes* (*HUCs*) are used by the U.S. Geological Survey to classify four levels of hydrologic units beginning with 21 major geographic regions that contain either major river basins or a series of river basins in a particular region. The major regions are subdivided into 221 subregions with 378 accounting units and ending with 2264 watershed units (Seaber et al., 1987). HUCs are used for mapping and describing areas on the landscape and in some cases coincide broadly with ecoregions (Omernik, 2003). HUCs can be watersheds or other land units that have similar characteristics of climate, vegetation, geology, soils, land use, and topography. As such, HUCs would be expected to have common hydrologic properties that differ from those HUCs with a different set of characteristics.

Methods of assessing watershed characteristics that are needed for certain applications are presented in the context of those applications in this book. A discussion of tools and emerging technologies for making hydrologic and watershed assessments is presented in Chapter 16.

Integrated Watershed Management

IWM involves an array of vegetative (nonstructural) and engineering (structural) practices (Gregersen et al., 2007). Soil conservation practices, constructing dams, and establishing protected reserves can be tools employed in IWM as can be land-use planning that entails developing regulations to guide timber-harvesting operations, road-building activities, urban development, and so forth. The unifying focus in all cases is placed on how these varying activities affect the relationships among land, water, and other natural resources on a watershed. The common denominator or the integrating factor is water. This focus on water and its interrelationship with land and other natural resources and their use is what distinguishes IWM from other natural resource management strategies.

On the one hand, IWM is an integrative way of thinking about people's activities on a watershed that have effects on, or are affected by, water. On the other hand, IWM includes tools or techniques such as the physical, regulatory, or economic means for responding to problems or potential problems involving the relationship between water and land uses. What sometimes confuses people is the fact that these tools and techniques are employed not only by those designated as "watershed managers" but also by foresters, farmers, soil conservation officers, engineers, and so forth. In reality, all are watershed managers.

This fact is both the dilemma and the strength of watershed management. In practice, activities using natural resources are decided on and undertaken by individuals, local governments, and various groups that control land in a political framework that has little relationship to, and often ignores, the boundaries of a watershed. For example, forested headwater areas of a river basin could be under the control of a governmental forestry agency but the middle elevations and lowlands could be a composite of private, municipal, and individual ownerships.

Activities are often undertaken independently with little regard to how they affect other areas. Despite this real world of disaggregated and independent political and economic actions, it remains a fact that water and its constituents flow from higher to lower elevations according to watershed boundaries, regardless of the political boundaries. What one person or group does upstream can affect the welfare of those downstream. Somehow, the physical facts of watersheds and the political realities have to be brought together to achieve IWM. Within this broad focus there is concern with both how to prevent deterioration of an existing sustainable and productive relationship between the use of land, water, and other natural resources and how to restore or create such a relationship where it has been damaged or destroyed in the past.

Embodied in IWM are

- preventive strategies aimed at preserving existing sustainable land-use practices; and
- restorative or rehabilitation strategies designed to overcome identified problems or improve conditions to a desirable level where *desirable* is defined in ecological, environmental, and political terms.

Both strategies respond to the same types of problems. However, in one case the objective is to prevent a problem from occurring while in the other case the objective is to improve conditions once the problem has occurred. In reality, we are dealing with a continuum from regulatory support and reinforcement of existing sustainable land-use practices (preventive strategy) to such actions as emergency relief, building of temporary water-control structures, restoring wetlands, or changing land use on fragile and eroded lands (rehabilitation strategy). These routine preventive strategies and actions are as important as the more dramatic and visible restorative actions. Losses avoided through preventive strategies are as important to people as gains from rehabilitating a degraded watershed. In economic terms, the cost of preventing losses of productivity in the first place can be much lower than the cost of achieving the same benefit through more dramatic actions to restore productivity on degraded lands.

Sustainable Use and Development of Natural Resources

There are many examples of the need to manage watersheds better to meet current and future demands for water and other natural resources in a more sustainable manner. It must be recognized, however, that practices relating to resource use and management do not depend solely on the physical and biological characteristics of watersheds. Institutional, economic, and social factors such as the cultural background of rural populations and the nature of governments need to be fully integrated into solutions that meet environmental, economic, and social objectives (Gregersen et al., 2007). How these factors are interrelated can best be illustrated by looking at specific issues.

Land and Water Scarcity

Arable land and water resources are becoming scarcer with the earth's expanding population of people. These scarcities and the human responses to these scarcities pose challenges to sustainable development and can have serious environmental consequences. Changing climatic conditions and weather patterns add uncertainty to future land and water resource management. It is unclear where and how much the supplies of freshwater will change with changes in climate and weather. What is clear, however, is that the increasing water demands caused by increasing populations of people and expanding economic developments will pose far greater problems by the year 2025 than currently (Vorosmarty et al., 2000).

Knowledge, information, and technologies will no doubt expand our capability to deal with shortages of resources. To harness these capabilities, however, will require an integrated and interdisciplinary approach to planning and managing natural resources. The lack of such an approach has been problematic as emphasized by Falkenmark (1997) who stated, "In environmental politics, land and water issues are still seen as belonging to different worlds, taken care of by different professions with distinctly different education and professional cultures." Land, water, and other natural resources have traditionally been managed in isolation of one another as a consequence. However, we recognize that land-use affects the quantity and quality of water flow in a watershed. Water development, conversely, affects land use. Understanding these relationships and linkages is essential to achieving sustainable use and development of natural resources.

Land Scarcity

Land scarcity has been exacerbated in many developing countries by the rural poor who clear forests to grow agricultural crops, cultivate steep uplands, and overgraze fragile rangelands to meet their food and natural resource needs. Watershed degradation frequently results, further reducing land productivity that in turn causes more extensive and intensive land use. In other instances, irrigation practices intended to expand agricultural productivity have been inappropriate and have caused land productivity to diminish through salinization. As watershed conditions deteriorate, the people living in both uplands and downstream areas become impacted. This cycle of deteriorating land productivity is a clear indicator of nonsustainable land use.

Of the 8.7 billion ha of agricultural land, forests, woodlands, and rangelands worldwide, almost 25% has been degraded since the mid-1900s with 3.5% being severely degraded

(Scherr and Yadav, 1996). Problems of desertification have received global attention as the productivity of some of the poorest countries in the world continues to decline. Ironically, we have the know-how to manage resources in a sustainable manner. What is most often lacking are the policies and institutions to promote and sustain sound resource use.

Water Scarcity

Water scarcity has attracted attention globally and is considered the major environmental issue facing the twenty-first century. The United Nations commemorated a World Day for Water in March 2001 where speakers concluded that demands for freshwater exceeded supplies by 15–20% and that two-thirds of the world's population will experience severe water shortages in the next 25 years. Although the 9000–14,000 km^3 of freshwater on earth should be sufficient to support expanding human populations for the foreseeable future, unequal distribution results in water scarcity (Rosegrant, 1997). For example, per capita freshwater supplies in Canada are approximately 120,000 m^3 compared to Jordan's 300 m^3. In China, per capita freshwater supplies are 2700 m^3 with 600 of its major cities suffering from water shortages. Although China encompasses the Yangtze River, the third largest river in the world, water scarcity is pronounced in areas north of the river which represent 63.5% of the country's land area but only 19% of its water resources. To enhance water supply in northern China, where more than one-third of the county's population lives, the largest water diversion project is underway to divert 44.8 billion m^3 of water annually from the Yangtze, Huaihe, and Haine rivers in the south to the drier north (www.water.technology.net/projects/south_north/). Other diversions are planned for western regions of the country.

Such programs have dramatic effects on land and water, placing greater importance on improved watershed planning and management to protect the life of major reservoir and diversion investments. The fact is that water has become a global environmental priority, exceeding climate change according to a GlobeScan/Circle of Blue Report (Box 1.2). Examples of the role of IWM in coping with water scarcity and other natural resource issues are outlined in Table 1.1.

UNESCO (www.unesco.org, 2011) estimates that global water withdrawals have tripled over the past 50 years. The International Water Management Institute (IWMI) paper of 2007 (www.iwmi.cgiar.org) reported that much of the increased use of water is attributed to irrigation of agricultural lands. Competition for water between agricultural and other interests (municipal and environmental) will undoubtedly increase in coming decades with the world's population projected to grow to 9 billion people by 2050.

Coping with Hydrometeorological Extremes

Floods, landslides, and debris torrents that result from excessive precipitation events and the droughts that result from deficient precipitation represent both dimensions of hydrometeorological extremes. The disasters and famine that can result present some of the greatest challenges to land and water resource managers. Even though we call them extreme events, it should be recognized that floods and droughts occur naturally and are not necessarily rare. However, forecasting when and where they will occur and at what magnitude is uncertain. Whether these hydrometeorological extremes end up as disasters depends on their impacts on people and natural ecosystems. What can be done to cope better with such extreme

Table 1.1. The role of watershed management in developing solutions to natural resource problems

Problem	Possible alternative solutions	Associated watershed management objectives
Deficient water supplies	Reservoir storage and water transport	Minimize sediment delivery to reservoir site; maintain watershed vegetative cover
	Water harvesting	Develop localized collection and storage facilities
	Vegetative manipulation; evapotranspiration reduction	Convert from deep-rooted to shallow-rooted species or from conifers to deciduous trees
	Cloud seeding	Maintain vegetative cover to minimize erosion
	Desalinization of ocean water	Not applicable
	Pumping of deep groundwater and irrigation	Management of recharge areas
Flooding	Reservoir storage	Minimize sediment delivery to reservoir site; maintain watershed vegetative cover
	Construct levees, channelization, etc.	Minimize sediment delivery to downstream channels
	Floodplain management	Zoning of lands to minimize human activities in flood-prone areas; minimize sedimentation of channels
	Revegetate disturbed and denuded areas	Plant and manage appropriate vegetative cover
Energy shortages	Utilize wood for fuel	Plant perpetual fast-growing tree species; maintain productivity of sites; minimize erosion
	Develop hydroelectric power project	Minimize sediment delivery to reservoirs and river channels; sustain water yield
Food shortages	Develop agroforestry	Maintain site productivity; minimize erosion; promote species compatible with soils and climate of area
	Increase cultivation	Restructure hillslopes and other areas susceptible to erosion; utilize contour plowing, terraces, etc.
	Increase livestock production	Develop herding–grazing systems for sustained yield and productivity
	Import food from outside watershed	Develop forest resources for pulp, wood, wildlife products, etc., to provide economic base
Erosion/sedimentation from denuded landscapes	Erosion control structures	Maintain life of structures by revegetation and management
	Contour terracing	Revegetate, mulch, stabilize slopes and institute land-use guidelines
	Revegetate	Establish, protect, and manage vegetative cover until site recovers

Table 1.1. (*Continued*)

Problem	Possible alternative solutions	Associated watershed management objectives
Poor-quality drinking water	Develop alternative supplies from wells and springs	Protect groundwater from contamination
	Treat water supplies	Filter through wetlands or upland forests
Polluted streams/ reduced fishery production	Control pollutants entering streams	Develop buffer strips along stream channels; maintain vegetative cover on watersheds; develop guidelines for riparian zones
	Treat wastewater	Use forests and wetlands as secondary treatment systems for wastewater

events is a critically important question that planners and managers as well as students of hydrology and watershed management should be prepared to address.

Floods, landslides, and debris torrents result in billions of dollars being spent globally each year for flood prevention, flood forecasting, and stabilization of hillslopes and stream channels. However, the cost of lives and property damage due to floods, landslides, and debris flows is staggering. The impacts of these naturally occurring phenomena can be exacerbated by human encroachment on floodplains and other hazardous areas, which is often the result of land scarcity.

Responses to disasters caused by too much water from flooding or too little water that results from droughts are most often short term and limited to helping those who have been directly impacted rather than dealing with the causes. Interest in coping with extreme events flourishes immediately after they have taken their toll on human lives and property but becomes lower priority with reduced funding as memories fade with time. The term *crisis management* captures this approach. It is not unreasonable to ask the question: what (if anything) can be done to prevent such disasters? There is no single approach that applies to all situations. To paraphrase Davies (1997), there are essentially three options available to reduce or prevent disasters caused by excessive amounts of water. These options are modifying the natural system, modifying the human system of behavior, and some combination of these two.

It is important to recognize that no matter how advanced our technologies might become, the degree to which such events result in disasters is largely because of the behavior of people on watersheds. Nevertheless, while solutions are not necessarily easily found, they must be based on sound hydrological principles.

Attempts to control events by modifying hydrologic and other natural systems most often entail the use of engineering measures including reservoirs, levees, and channelization aimed at controlling floods (see Table 1.1). There are increasing attempts to apply vegetative management and bioengineering measures along with structures to exert some control over extreme events. However, it is unrealistic to believe that we can mitigate the most extreme of hydrometeorological events. The development of such measures requires an understanding of natural hydrologic systems and processes, how these systems function, and how mitigation impacts other aspects of the environment.

Box 1.2

Water – The Top Global Environmental Concern?
http://www.circleofblue.org/waternews/2009/world/waterviews-water-tops-climate-change-as-global-priority/. Accessed April 7, 2010

A poll of 1000 people in each of 15 countries, including Canada, China, India, Mexico, Russia, United Kingdom, and the United States, suggested that fresh clean water has a greater impact on people's quality of life, exceeding concerns of air pollution, species extinction, loss of habitat, depletion of natural resources, and climate change. Water concerns range across the spectrum of water resources with top concerns expressed by countries as follows:

- India – lack of safe drinking water; water pollution; access to fresh clean water is highest priority; the Yamuna River is one of the most polluted waterways in the world.
- Russia – industrial water pollution of major rivers is of foremost concern.
- China – industrial pollution of the Yellow River and water scarcity plague many areas of China.
- United States – dependence on the Colorado River for irrigation in Imperial Valley and Coachella Valley of southern California has made water extremely scarce and valuable; the demise of the Salton Sea is attributed to high consumptive use and evaporation of water plus pollution from agricultural chemicals and trace metals.
- Mexico – Mexico City's supply is insufficient to meet the demands of its human population of more than 25 million; this has led to the depletion of underground aquifers and impaired waters due to human sewage.

We can view the options presented in Table 1.1 as those that attempt to move or control water and those that attempt to move or manage people on a watershed. Furthermore, some of the measures listed tend to be the responsibility of water resource organizations with an engineering bias such as governmental agencies responsible for flood control or irrigation projects, or private utilities companies responsible for hydropower. Organizations responsible for land management often have a land production–ecological viewpoint. Too often the separation of responsibilities and the lack of coordination lead to piecemeal programs that result in unwanted effects and, furthermore, do not have the long-term perspective needed to develop sustainable solutions.

While it is true that land use might not affect the magnitude of the most extreme hydrometeorological events, people's activities on a watershed and the cumulative effects of these activities determine the extent to which the events impact human life and welfare.

The reliance on large-scale engineering structures to store or divert water with the intention of meeting water management objectives has been met with objections from environmental groups for some time. However, there is concern by some professionals that an over-reliance on engineering measures such as dams, levees, or stream channelization has

- led to unintended and unwanted effects;
- reduced the hydrologic function and environmental values associated with natural rivers, floodplains, and estuaries; and
- imparted a false sense of security to those living in downstream areas.

Agricultural expansion and urban development have further reduced natural riparian communities, wetland ecosystems, and floodplains with often serious consequences (Box 1.3).

Box 1.3

Effects of Altering Natural Ecosystems on Flooding: Examples in the Upper Midwest, United States

The aftermath of record flooding in the upper Mississippi River in 1993 and 2001, and the 1997, 2001, 2007, and 2010 floods in the Red River of the North along the Minnesota–North Dakota boundary, led to questions concerning the extent to which the floods were the result of human modifications of watersheds. Since the beginning of the twentieth century, efforts designed largely to expand farmland have resulted in extensive drainage of wetlands and the use of tile drains on croplands. Flood control has involved levee construction and channelized rivers, with a resulting loss of floodplain storage and riparian forests.

Following the 1993 Mississippi River flood that caused $10 billion in damage (Tobin and Montz, 1994), Leopold (1994) found that in several locations the levees caused river stages to be higher than they would have been without levees being in place. As long as levees did not fail or were breached, the increased height of flood stages was retained within the channel system. However, many levees were breached, causing greater devastation than would have otherwise occurred without levees. The cumulative loss of wetland and floodplain storage along channels and in upper watersheds could have exacerbated flooding in many locations. The question of the extent to which land use and channel modifications have affected the magnitude of floods throughout the Mississippi River and the Red River of the North is uncertain. A better understanding of cause-and-effect relationships is needed; this book provides the background to help understand and address these issues.

Watersheds, Ecosystem Management, and Cumulative Effects

Taking a watershed perspective has utility in addressing the many natural resource issues that span decades and involve varying climatic and land-use changes. The merit in taking a watershed perspective is clear when dealing with issues of water supply and events such as floods and droughts (see Table 1.1). Not so obvious are the effects that become compounded spatially and temporally and that at first glance might not appear to be related to hydrologic processes.

During the past century in the United States, vegetation and land-use changes on the landscape have dramatically altered the hydrologic characteristics of watersheds. Since the earliest European settlers, extensive areas of native forests and grasslands have been converted to agricultural cropland. Urban areas and road systems have expanded, riparian corridors altered, wetlands drained, and natural river systems modified. These landscape changes have resulted in hydrologic changes through modifications of watersheds, their stream systems, and surface–groundwater linkages. Changes in water flows and water quality can affect people and ecosystems in both upstream and downstream areas.

The challenge to people living on watersheds is that of mitigating the effects of all of the changes that adversely impact human welfare and the functioning of natural ecosystems. Increased attention is being paid to maintaining or restoring natural stream-channel systems, riparian communities, wetland ecosystems, and floodplains as needed to maintain the watersheds in a good hydrologic condition. For example, Hey (2001) called for a major program to maximize the natural storage of wetlands and floodplains and minimize conveyance in the upper Mississippi River Basin. Such a program would in effect reverse some of the effects of the past 200 years of levee construction and other engineering practices in the basin.

The role of watersheds as units for ecosystem management and analysis has gained international recognition in the past decades. *Ecosystem management*, a focus of natural resource management in the 1990s, is based on maintaining the integrity of ecosystems while sustaining benefits to human populations. The usefulness of a watershed approach in ecosystem management became apparent in the United States with the attempts to reconcile conflicts over the use and management of land and water in the Pacific Northwest (Montgomery et al., 1995). Environmental, economic, and political conflicts involving the use of old-growth forests for both spotted owl habitats and the production of timber were heated in this region. At the same time, concern over the effects of timber harvesting and hydropower dams on native salmon emerged. A watershed analysis approach facilitated the examination of the interrelationships between land and water use and many of the environmental effects. This approach was taken in the Pacific Northwest because watersheds define "basic, ecologically, and geomorphologically relevant management units" and "watershed analysis provides a practical analytical framework for spatially-explicit, process oriented scientific assessment that provides information relevant to guiding management decisions" (Montgomery et al., 1995, p. 371).

Cumulative watershed effects are the combined environmental effects of activities in a watershed that can adversely impact beneficial uses of the land (Reid, 1993; Sidle, 2000). The viewpoint taken here is that there are interactions among different land-use activities and there can be incremental effects that can lead to more serious overall impacts when added to past effects. Individually, these environmental effects might not appear to be relevant but collectively they can become significant over time and space. For example, the conversion from forest to agricultural cropland on one part of a watershed can cause an increase in water and sediment flow. Road construction and drainage can have similar effects elsewhere in a watershed as can the drainage of a wetland at another location.

These activities are likely to occur over a long period and, incrementally, have little obvious effect. At some point in time, however, the increased streamflow discharge in combination with additional sediment loads can lead to more frequent flooding. River channels can adjust to these changes in the flow of water and sediment to cause additional impacts downstream. These same changes can alter aquatic habitats. A watershed perspective forces one to look at multiple and cumulative effects in a unit and attempt to identify key linkages between terrestrial and aquatic ecosystems.

Reconciling Watershed and Political Boundaries

Historically, land, water, and other natural resources have been managed not only by people with different technical backgrounds but also by organizations that have very focused and different missions that impact one another. There are often overlapping responsibilities. In contrast, it is rare to find organizations that have the explicit mission of watershed management. There are few organizations with responsibilities that coincide with watershed boundaries as a result. However, there is a need to cope with issues that arise between upstream and downstream entities. In response to these needs, watershed management organizations and institutions have emerged with responsibilities of planning and managing resources to resolve upstream–downstream conflicts that arise (see NRC, 2000).

Locally led watershed management organizations and initiatives have been established under various names, including watershed associations, partnerships, councils, and other “co-management” schemes administered jointly by governments and local communities. They all emphasize the growing awareness of the important hydrologic linkages between uplands and downstream areas. Furthermore, there is a growing recognition that institutional arrangements and supportive policies are needed so that people can develop sustainable solutions to land and water problems and avoid land and water degradation. By the late 1990s, more than 1500 locally led initiatives were established in the United States in response to water resource problems that were not being addressed (Lant, 1999). These initiatives were largely in response to the issues discussed above, including the effects of urbanization, intensification of agriculture, forest management activities, stream channelization, and wetland drainage on flooding, water yield, and water quality.

A river basin management strategy has emerged in North Carolina, United States (www.ncwater.org/basins, accessed May 24, 2011), as a result of a decade of water supply demands exceeding water supplies in many locations of the state. Water shortages during a 4-year drought intensified in 2002, suggesting that actions would be needed to avoid adverse economic impacts caused by insufficient water supplies. The river basin strategy took the approach that river basins provide the fundamental units for local and state governmental organizations to monitor water availability and use and to facilitate planning to develop 50-year water supply plans to solve water supply problems. A similar strategy, but for

Box 1.4

Minnesota's Major Watershed Restoration and Protection Strategy
In November of 2008 after the US stock market crashed and votes throughout the United States sent Barack Obama to Washington DC, Minnesotans voted to raise their sales tax by 3/8's of a percent. A new clean water and land legacy fund was created to clean up the state's impaired waters and preserve valued hunting land and wetlands over a 25-year period. The state developed a strategy for not only restoring impaired waters, but also identifying land at risk of developmental changes that could adversely impact water resources in the future. The strategy focused on several key areas:

(1) Assess the overall condition and health of the many lakes, streams, rivers, wetlands by major watersheds using biological metrics in the form of an Index of Biological Integrity (IBI). A major watershed is defined as an eight-digit HUC according to the U.S. Geological Survey; Minnesota has 81.
(2) Develop an assessment document, including additional data such as Lake Secchi, Chlorophyll A, and total Phosphorus that would guide listings for the Section 303(d) list of the Clean Water Act.
(3) Conduct a Stressor Identification analysis to validate true water quality impairment. This step involves more than examining toxic influences upon fish; hydrologic, geologic, geomorphic, and connectivity data must be gathered to define natural background and superimposed anthropogenic activity.
(4) Construct computer models for each major watershed to help manage current land-use conditions and predict future water quality scenarios based on climate and land-use change.
(5) Perform statistical analysis and synthesis of all data and model results to identify Priority Management Zones for restoration or protection action.
(6) Lastly, with Clean Water Legacy funds implement Best Management Practices where they are most needed on the landscape to form watershed-based treatment trains from upland through riparian and into channels, lakes, and wetlands.

This approach is unique in the United States; no other state has this level of local funding focused on an IWM approach (www.pca.state.mn.us/index.php/view-document.html?gid = 14224).

different purposes, has been developed in Minnesota where major watersheds are used as the units of management to meet water quality objectives (Box 1.4).

Institutional responses to international water issues are also emerging. For example, transboundary issues in the Nile River Basin have focused attention on watersheds as a logical framework for analysis and understanding of the effects of human actions in one part of the basin on those living downstream. Here, concerns about water, human poverty, biodiversity, and wildlife are paramount. The disparity between watershed and political boundaries prompted the Nile Basin Initiative and an extensive study into transboundary issues of the Nile River Basin (Box 1.5). The inequities in rainfall and water supply among the ten riparian countries of the Nile River Basin have threatened sustainable development in the region and have historically resulted in political conflict (Baecher et al., 2000). While a necessary and daunting task, developing solutions to water supply problems is

Box 1.5

The Nile River Basin: A Case for Watershed Management (Baecher et al., 2000)

The Nile River is the longest river of the world, traversing 6700 km from the rift valleys of East Africa, connecting with the Blue Nile from the Ethiopian highlands, and emptying into the Mediterranean Sea through the broad delta of Egypt. Annual rainfall amounts vary to the extreme within the basin. The Ethiopian highland watersheds experience more than 2000 mm of annual rainfall and contribute 60–80% of the water flow in the Nile River, yet represent less than 10% of the land area in the Nile Basin. Vast areas of the northern Basin, on the other hand, average less than 50 mm of annual rainfall. Consequently, the watersheds of Egypt and Sudan, which constitute more than 75% of the basin, contribute negligible flow to the Nile River.

Efforts to relocate waters of the Nile Basin for economic development have led to environmental concerns and potential political conflict. Lake Nassar, created by the large Aswan Dam of the lower Nile River, is essential to Egypt's economy but is vulnerable to water losses upstream. The controversial Jonglei Canal was proposed to enhance downstream water supplies by diverting water from the vast Sudd wetland to the lower and drier north – at the expense of the loss of valuable habitat. In contrast, Ethiopia as one of the poorest countries in the basin would benefit economically by expanding its irrigation and hydropower production through construction of dams in the uplands. However, such projects would meet with objections from downstream riparian countries. Furthermore, such projects, whether in uplands or lowlands, are threatened by intensive land use, watershed degradation, and deforestation that impair the quantity and quality of streamflow and aquatic habitat. Forest cover in the Ethiopian Highlands decreased from more than 15% in the 1950s to less than 4% (43.4 million ha) by 2000. FAO (2007) indicated that annual deforestation was occurring at a rate of 1410 km^2/year. Overgrazing and cultivation of hillslopes lead to soil erosion, loss of productivity of the land, and increased sediment delivery downstream. All these issues focus on the need for watershed management that extends beyond country boundaries.

not necessarily sufficient. The growth and distribution of human populations, widespread poverty, and inadequate policy responses to land and water resource issues compound problems in the basin. A major theme in the Nile Basin Initiative is that land and water must be managed in harmony so that environmental conditions and human welfare benefit. IWM is one of the key themes in this effort.

A proactive approach to IWM involves establishing and sustaining preventive practices. This approach requires instituting guidelines of land use and implementing land-use practices on a day-to-day basis that result in long-term, sustainable resource development and productivity without causing soil and water problems. Land and natural resource management agencies in the United States and many other countries have established policies that embody sound watershed management principles. These preventive measures might not make for exciting reading but in total they represent the ultimate goal of IWM. Achieving this goal requires that managers recognize the implications of their actions and that they work effectively within the social and political setting in which they find themselves.

The dilemma in watershed management is that land-use changes needed to promote the survival of society in the long term can be at cross-purposes with what is essential to the survival of the individual in the short term. Requirements for food, high-quality water, and the commodities and amenities provided by natural resources today should not be met at the expense of future generations. Therefore, any discussion of sustainable natural resource development should consider watershed boundaries, the linkages between uplands and downstream areas, and the effects of land-use practices on long-term productivity. Land use that is at cross-purposes with environmental capabilities cannot be sustained. However, sustainable productivity and environmental protection can be achieved with the integrated, holistic approach that explicitly considers hydrology and the management of watersheds.

Summary and Learning Points

Watersheds are biophysical systems that describe how units of land are connected by water flow. They also represent systems that are suitable for developing sustainable use and management of multiple resources under the tenet that land and water should be managed in concert with one another – the basis for IWM. This introductory chapter lays out the reasons for taking an IWM approach – which the reader should be able to understand – as well as the role of IWM in coping with land and water scarcity and hydrometeorological extremes.

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CHAPTER 2 Hydrologic Cycle and the Water Budget

INTRODUCTION

Hydrology is the science and study of water. In this book, we focus on the occurrence of water on earth and the factors that influence the amount, distribution, circulation, timing, and quality of water from a watershed perspective. Water flows in accordance with physical laws and in many ways alters the landscape features of watersheds. In conjunction with climate, water affects the development of soils and the type of vegetative cover that grows on a watershed. The quantity, quality, and timing of water that flows from watersheds are dependent on the interactions of climate and the activities of people on the watershed. An understanding of the causes, processes, and mechanisms of water flow on earth and through its watersheds provides the underpinning for the study of watershed hydrology.

This chapter begins with a review of the properties of water and examines the basic principles and processes involved in the flow of water from a watershed and the corresponding flows of energy that drive the hydrologic cycle.

PROPERTIES OF WATER

Water is commonplace but also unique. As the most abundant compound on the earth's surface, water is the only common substance that exists naturally in all three states – liquid, solid, and gas. The chemical formula for water, H_2O, is two hydrogen atoms covalently bonded to one oxygen atom. The water molecule is polar; that is, it has a net positive charge on the hydrogen atoms and a net negative charge on the oxygen atom because hydrogen attracts electrons less strongly than oxygen. This property of polarity results in the physical attraction of water molecules to one another (hydrogen bonding) that explains many of the physical and chemical properties of water that are integral to life on earth.

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

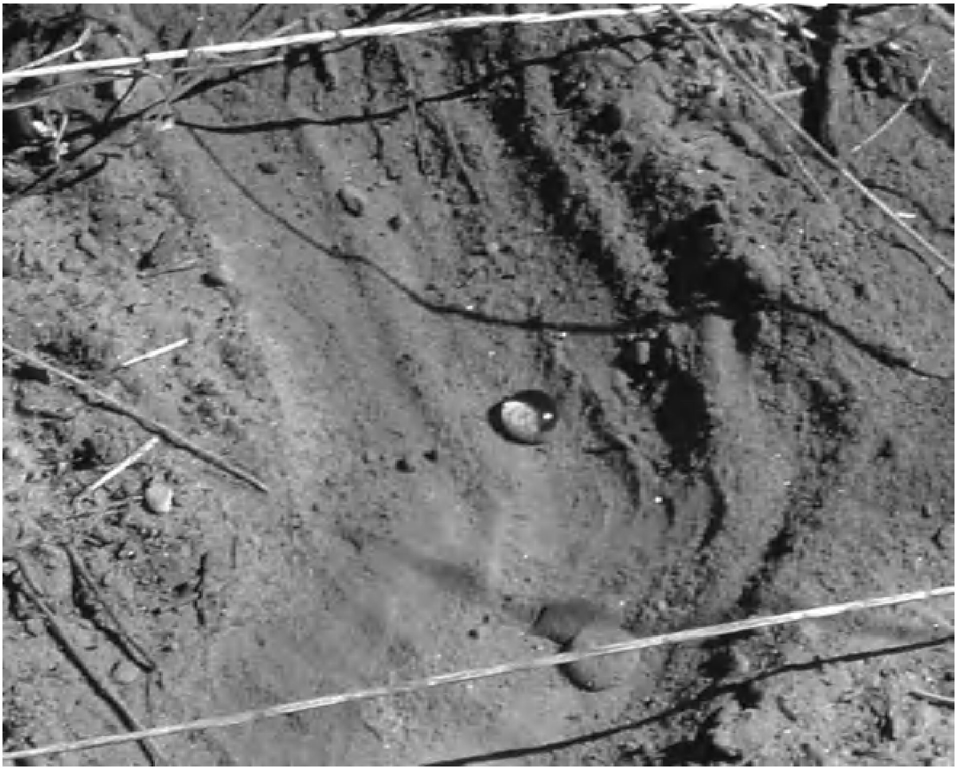

FIGURE 2.1. Water droplet on hydrophobic soil illustrates the cohesive property of water molecules

Importance of Polarity

Water polarity results in the important properties of cohesion, adhesion, and capillarity. Hydrogen bonds help keep water molecules together (cohesion) while polar bonding to other materials (adhesion) explains the attraction of water molecules to soil particles and plant cells. The cohesion between water molecules explains the high surface tension of water, which is the reason why a water droplet maintains its form as a bubble on a nonabsorbent or water-repellent surface (Fig. 2.1). The properties of surface tension and adhesion explain why water can move upward into a small capillary tube against the force of gravity. This capillary action, or capillarity, also explains why water can move against the force of gravity through soils and plants. Water moves into and up a plant as a result of both the cohesion of water molecules that keep the water column intact and the adhesion properties of water that attach water to plant cell walls in the xylem. The cohesive properties and surface tension of water molecules explain why water droplets form in the atmosphere and grow in size until they fall to the earth's surface under the influence of gravity.

The polarity of water molecules results in water being a good solvent, sometimes referred to as a universal solvent. While not all substances such as oils and waxes mix with water, many substances such as other polar or ionic molecules (salts, acids, and alcohols) mix well with water. As a result, water plays an important biological role in transforming

Table 2.1. The unique water molecule properties of water at normal terrestrial temperatures (Maidment 1993 and others)

Temperature	Density		Heat capacity (C_p)		Heat of vaporization (L)	
(°C)	(kg/m^3)	(g/cm^3)	(J(g/K))	(cal/(g/°C))	(J/g)	(cal/g)
0 (ice)	916.70	0.92	2.05	0.5	2835	677.3
0 (liquid)	999.84	1.00	4.22	1.0	2500	597.3
4	999.97[a]	1.00	4.19	1.0	2491	595.2
10	999.70	1.00	4.18	1.0	2477	591.7
20	998.21	1.00	4.18		2453	586.0
30	995.65	1.00	4.18		2429	580.4
40	992.20	0.99			2406	574.9

[a] Maximum density of water and an important characteristic that explains why the water in lakes becomes stratified in summer and turns over as water cools in the fall/early winter.

and moving many mineral salts and nutrients in solution through soils to plants and from groundwater. Any nutrients and mineral salts that are in solution in soil water are easily moved by excess water draining a soil or flowing off the soil surface and subsequently entering groundwater or surface water bodies. Because of the physical forces of water flowing across the soil surface, both nonsoluble and soluble materials can be transported into stream channels and other water bodies.

State of Water

The form or state of water that occurs at any location on earth, whether gas, liquid, or solid, is a function of the properties of water and energy. Most water on earth occurs in liquid form. Liquid water has the highest *specific heat capacity* of all known substances (Table 2.1). Changes in state from liquid to water vapor or to ice and vice versa involve exchanges of energy. Large amounts of energy are required to convert liquid water to its gaseous state, called the *latent heat of vaporization*. Therefore, the transformation of liquid water to water vapor, called *evaporation*, consumes large amounts of the total solar energy absorbed by the earth's surface.

The energy required to change ice at 0°C to vapor requires the sum of the heat of fusion (the energy required to melt ice is 334 J/g or 80 cal/g at 0°C) plus the latent heat of vaporization for liquid water at 0°C. Given the abundance of water on earth, these properties of specific heat and latent heat of vaporization explain why water acts as a thermal regulator that moderates the earth's climate.

In contrast to evaporation, the transformation of water vapor in the atmosphere to liquid water, a process called *condensation*, releases the latent heat of vaporization. When condensation occurs, energy becomes available to heat the air, melt snow, or heat the surface on which condensation occurs. The changes in state of water in the atmosphere affect air temperature, precipitation (see Chapter 3), and weather conditions in general.

Water vapor mixes completely in air and exerts a vapor partial pressure, called *vapor pressure*. The *saturation vapor pressure* (e_s) is determined by air temperature alone and is the partial pressure of water vapor (*mb*) in a saturated atmosphere as described by Lee (1978):

$$\ln e_s = 21.382 - \left(\frac{5347.5}{T}\right) \quad \textbf{(2.1)}$$

where T is the absolute temperature in K, which is equal to °C + 273.

At any given temperature, a parcel of air in an unsaturated atmosphere will have a partial pressure that is less than the saturation vapor pressure: the drier the air, the greater the difference between the vapor pressure of the air and its saturation vapor pressure. The magnitude of this vapor pressure difference affects the rate and amount of evaporation (discussed in Chapter 4).

Atmospheric pressure (not to be confused with vapor pressure) decreases with height, causing a rising parcel of air to expand. As a result of this expansion, the temperature of the air will decrease while that of a descending parcel of air will become compressed and warmed. These relationships can be explained by the natural or ideal gas law:

$$PV = nRT \quad \textbf{(2.2)}$$

where P is the pressure in pascals; V is the volume in m^3; n is the number of moles of gas; R is the universal or ideal gas constant and has the value 8.314 J/K/mol; and T is the air temperature in K.

The process is called *adiabatic* if no heat is gained or lost in the parcel of rising or falling air mass by mixing with surrounding air. The rate at which unsaturated air cools by adiabatic lifting is approximately 1°C/(100 m), called the *dry adiabatic lapse rate*. Movements of large air masses can change the lapse rate and conditions can even reverse the lapse rate when large cold air masses underlie less-dense warm air, a condition called an *inversion* in which temperature increases with height.

The properties of water and the interactions of water and energy provide the foundation for understanding the circulation of water and energy across the globe.

THE HYDROLOGIC CYCLE

The circulation of water on earth, called the *hydrologic cycle*, involves the processes and pathways by which water evaporates from the earth's surface to the atmosphere and returns to the surface as precipitation or condensation (Fig. 2.2). With the earth's surface being about 70% water, most of the atmospheric water originates from the oceans and other water bodies. With few exceptions, much of the precipitation falling on land surfaces does not reach the oceans as streamflow or groundwater flow but rather evaporates back into the atmosphere. This partitioning of water between evaporation, storage, and watershed liquid water flows involves numerous hydrologic processes.

Hydrologic Processes

Water evaporates from many surfaces located throughout a watershed. Precipitation that is caught by plant surfaces and evaporates back to the atmosphere is called *interception*. This part of precipitation does not reach the soil surface. Evaporation also occurs from water bodies located in a watershed and from soil surfaces. Water that is extracted from the soil by plant roots and that evaporates from within plant leaves is *transpiration*. The total amount of water that evaporates from a watershed is *evapotranspiration*, which is the sum

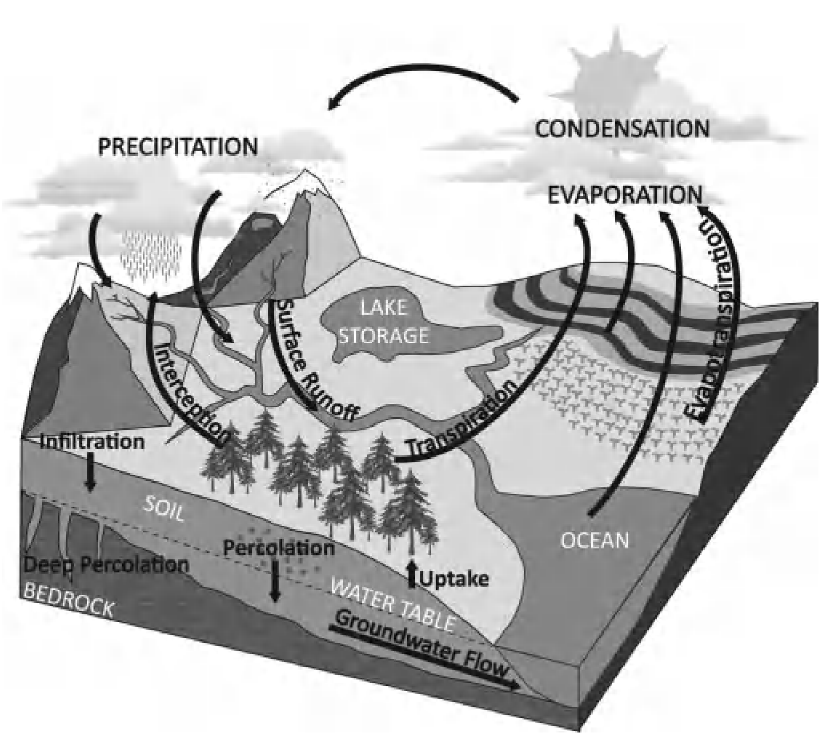

FIGURE 2.2. The hydrologic cycle (For a color version of this figure, see the color plate section)

of interception, transpiration, and evaporation from soils and water bodies. This evaporated water is temporarily lost from the watershed to the atmosphere but eventually returns to the earth's surface as precipitation at some other location and the cycle continues.

Precipitation falling on a watershed that is not returned to the atmosphere via evapotranspiration can either flow over the soil surface, reaching stream channels as overland flow, or *surface runoff*, or it infiltrates into the soil. The fate of infiltrated water depends on

- the moisture status of the soil;
- the water-holding capacity of the soil; and
- the network and size of pores within the soil matrix.

Infiltrated water that is in excess of the soil water holding capacity can flow downward under the influence of gravity until reaching groundwater. If the downward drainage of water, called *percolation*, reaches a strata of soil or rock with limited *permeability*, water can be diverted laterally through the soil and discharge into a stream channel or other surface water body as *subsurface flow*, often called *interflow*. The fate of water that reaches groundwater depends on subterranean characteristics of earth materials and geologic strata that influence the pathways by which groundwater can flow. Some groundwater intersects river channels or other water bodies, thereby returning to surface waters. Water that seeps into deep groundwater *aquifers* can be stored for centuries before returning to surface

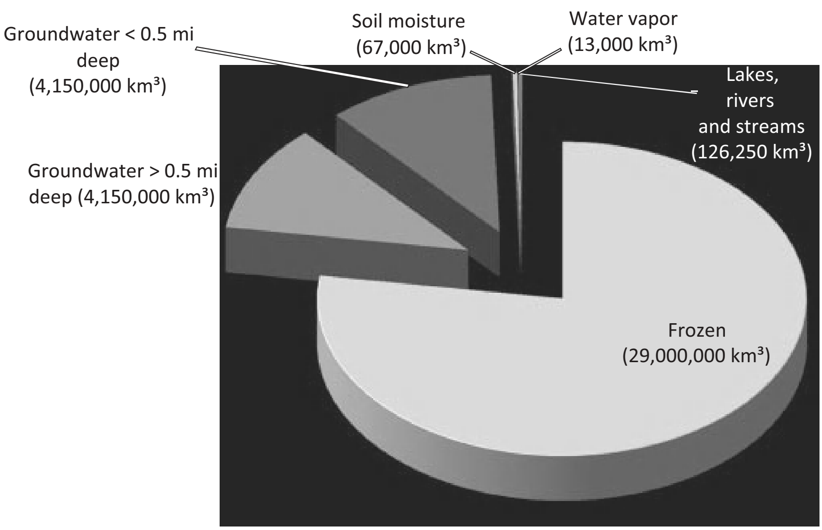

FIGURE 2.3. Distribution of freshwater on earth from van der Leeden et al. (1990)

waters. Therefore, the journey of water falling as precipitation on watersheds can follow a myriad of pathways to the ocean only to be evaporated and recycled again.

A fundamental concept of the hydrologic cycle is that water is neither lost nor gained from the earth over time. However, the quantities of water in the atmosphere, soils, groundwater, surface water, glaciers, and other components are constantly changing because of the dynamic nature of the hydrologic cycle. A few facts to remember about the hydrologic cycle are as follows:

- Solar energy provides the energy that drives and sustains the cycling of water on earth.
- There is no beginning or end to the cycle.
- The supply of water on earth is constant, but the allocation of water in storage or in circulation can vary with time.

The portion of water that is in various types of storage can be approximated over periods of time. If we consider the total water resource on the earth, only about 2.7% is freshwater of which about 77% exists in polar ice caps and glaciers (Fig. 2.3). About 11% of freshwater is stored in deep groundwater aquifers leaving about 12% for active circulation. Of this 12%, only 0.56% exists in the atmosphere and in the biosphere. The biosphere is from the top of trees to the deepest roots. The atmosphere redistributes evaporated water by precipitation and condensation. Components of the biosphere partition this water into runoff, soil and groundwater storage, groundwater seepage, or evapotranspiration back to the atmosphere.

The hydrologic processes of the biosphere and their interactions with vegetation and soils are of interest in understanding the hydrology and management of watersheds. This aspect of hydrology has been long emphasized in forest hydrology and watershed management courses taught in many natural resource programs; the emergence of *ecohydrology* in many academic programs takes this same perspective. Precipitation and the flow of water into, through, and out of a watershed all can be affected by land use and management activities. Likewise, human activities can alter the magnitude of various watershed storage components including soil water, snowpacks, rivers, lakes, and reservoirs. With the water budget approach, we can examine existing watershed systems, quantify the effects of management impacts on the hydrologic cycle, and in some cases predict or estimate the hydrologic consequences of proposed activities.

Water Budget

The hydrologic cycle is complex and dynamic but can be simplified if we categorize components into input, output, and storage (Fig. 2.4). Over any period of time, all inputs of water to a watershed must balance with outputs from the watershed and changes in storage of water within the watershed. This hydrologic balance, or *water budget*, follows the principle of the conservation of mass law expressed by the equation of continuity:

$$I - O = \Delta S \tag{2.3}$$

where I is the inflow, O is the outflow, and ΔS is the change in storage.

If inflows exceed outflows over a period of time, storage must increase, or if outflows exceed inflows then, storage must decrease – it is the conservation of mass law. The water budget can be expanded to identify the key inputs and outputs for a watershed over a specific period of time as follows:

$$P + GW_{\mathrm{i}} - Q - ET - GW_{\mathrm{o}} = \Delta S \tag{2.4}$$

where P is the precipitation (mm); GW_{i} is the groundwater flow into the watershed (mm); Q is the streamflow from the watershed (mm); ET is the evapotranspiration (mm); GW_{o} is the groundwater flow out of the watershed; ΔS is the change in the amount of storage in the watershed, $S_2 - S_1$ (mm), where S_2 is the storage at the end of a period and S_1 is the storage at the beginning of a period.

Note that in the water budget of a watershed (Equation 2.4), groundwater flows only need to be taken into account if they contribute to, or diminish, surface water on the watershed. Therefore, deep groundwater exchanges that have no contact with surface water are not considered. In some situations, there can be surface water seepage to deep groundwater aquifers, which is a net loss of water from the watershed, expressed as GW_{o}.

Soil Moisture and the Water Budget

The status of soil moisture in first-order watersheds largely dictates what portion of rainfall or snowmelt contributes to streamflow or to groundwater recharge. Therefore, the status of soil water content at any point in time, referred to as *antecedent soil moisture*, is an important hydrologic condition of a watershed that determines how the watershed responds. The change in storage for most first-order watersheds is determined largely by changes in soil moisture that is in response to precipitation, ET, runoff or streamflow (Q), and groundwater leakage (if any) over a period of time.

The amount of water that a soil can store is largely determined by soil texture (Fig. 2.5), although organic material in topsoil can increase the water-holding characteristics associated with the soil texture. *Field capacity* (FC) refers to the maximum amount of water that a given soil can retain against the force of gravity. If a soil were saturated and allowed

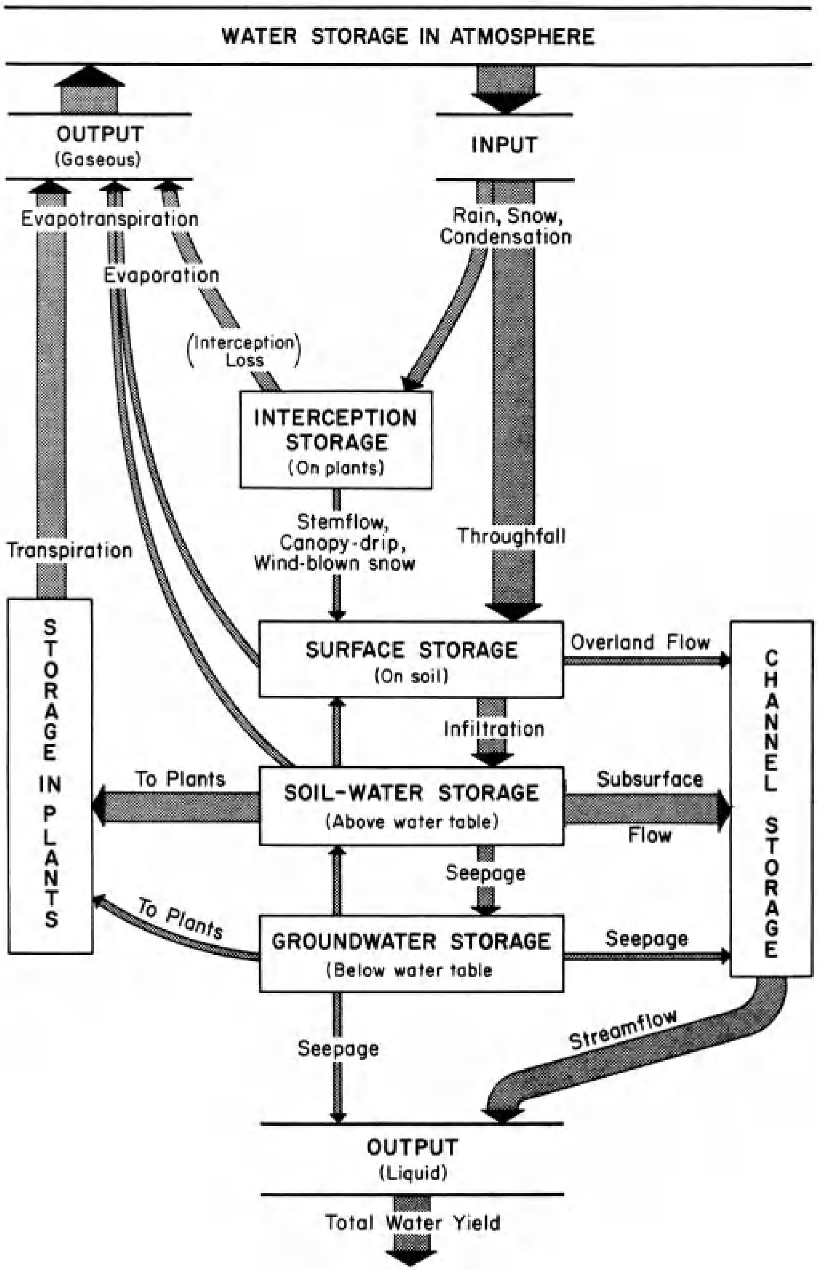

FIGURE 2.4. The hydrologic cycle consists of a system of water storage compartments and the solid, liquid, or gaseous flows of water within and between the storage points (from Anderson et al., 1976)

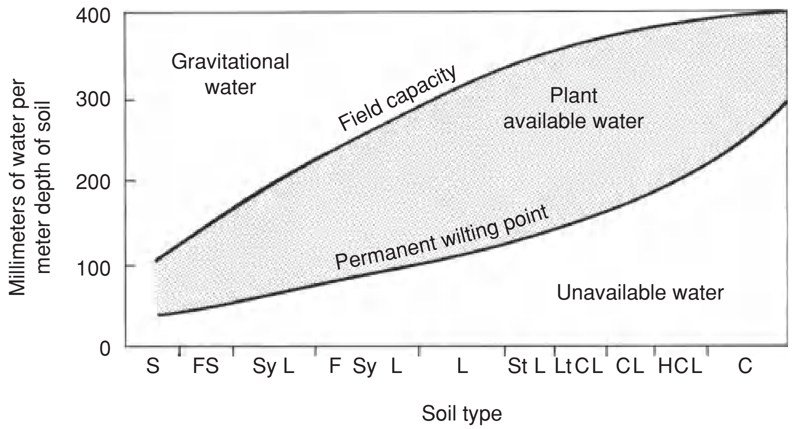

FIGURE 2.5. Typical water-holding characteristics of soils of different textures (redrawn from U.S. Forest Service, 1961). C, clay; F, fine; H, heavy; L, loam; Lt, light; S, sand; St, silt; and Sy, sandy

to drain freely, the amount of water remaining in the soil after all drainage ceased would be its FC. Evaporation at the soil surface and uptake of soil water by plant roots between precipitation events will dry the soil. If excessive drying of the soil occurs, the *permanent wilting point* (*PWP*) eventually can be reached, a point at which most plants can no longer survive because of limited soil moisture. The upper curve in Figure 2.5 is the *FC* relationship and the bottom curve is the *PWP* for different soil textures. The soil moisture content between *FC* and *PWP* represents plant available water. The depth at which plant roots grow in the soil determines how much of the plant available water can be extracted through transpiration.

Soil water above the *FC* drains freely through the soil unless drainage is impeded by some constraining layer. A soil can become saturated temporarily if the rate of infiltrated rainfall or snowmelt into the soil exceeds the rate at which water moves downward by *percolation* through the soil. For example, if the soil has an impeding layer with limited permeability, then the soil can become saturated above the impeding layer following excessive infiltration. Where surface runoff frequently collects, such as in landscape depressions, fine soil materials can settle and organic materials can accumulate forming an impeding layer at the bottom of the pond. When saturation conditions persist, an anaerobic environment forms above the impeding layer that can result in the formation of wetland ecosystems.

Soil moisture content measurements are often needed to determine change in soil moisture storage for water budget calculations. Soil moisture can be estimated gravimetrically (Box 2.1), with a neutron attenuation probe, or by dielectric constant methods (time domain reflectometry and frequency domain reflectometry).

For water budget applications, soil moisture content needs to be expressed as a volume. As shown in Figure 2.6, the amount of water in the void space of a 130-cm column of soil can be expressed as an equivalent depth of water, in this case 15 cm. The method of determining soil moisture volumes for use in water budget calculations is shown in Box 2.1. In developing water budgets for watersheds, it is convenient to express values of

Box 2.1

Steps in Determining Soil Water Content and Change in Soil Moisture for Water Budget Application

1. Determining soil moisture content by the gravimetric method:
   - Collect a sample of soil from the field and place in watertight container (in this case, the volume of soil sample = 20 cm^3)
   - Weigh the field soil sample = 31.8 g
   - After drying the soil in an oven to remove all water, the dry soil weight = 25.4 g
   - The water content by weight = (field weight − dried weight)/dried weight = (31.8 g − 25.4 g)/25.4 g = 0.25% or 25%
2. Determining soil moisture content by volume:
   - Determine the bulk density of the soil which is defined as the weight of dry soil per volume; in this case, bulk density = 25.4 g/20 cm^3 = 1.27 g/cm^3
   - The water content by volume can be determined directly as

   $$\begin{aligned} &= \text{(wet weight} - \text{dry weight)/sample volume} \\ &= (31.8\ \text{g} - 25.4\ \text{g})/20\ \text{cm}^3 \\ &= 6.4\ \text{g}/20\ \text{cm}^3 \\ &= 0.32\ \text{g/cm}^3 \end{aligned}$$

   since 1 g of water = 1 cm^3, the water content by volume is 32%, or it can be determined from the bulk density measurement as
   bulk density × water content by weight = 1.27 g/cm^3 × 0.25
   = 0.318% or 32%
3. Determining soil moisture content for a column of soil:
   If this soil moisture content was representative of a soil that was occupied by plant roots to a depth of 130 cm deep, then the total amount of water in that soil profile would be

   $$130\,\text{cm} \times 0.32 = 41.6\,\text{cm}$$

4. Determining change in soil moisture storage over time:
   If the preceding sample of soil has a moisture content of 41.6 cm at June 1 and the soil was again sampled on June 30 and found to have a moisture content of 37.5 cm, then the change in storage (ΔS) over the month of June would be

   $$\Delta S = S_2 - S_1 = 37.5\,\text{cm} - 41.6\,\text{cm} = -4.1\,\text{cm}$$

   where S_2 is the soil moisture at the end of the period (June 30) and S_1 is the soil moisture at the beginning of the period (June 1). Thus, the soil lost 4.1 cm of moisture over the period.

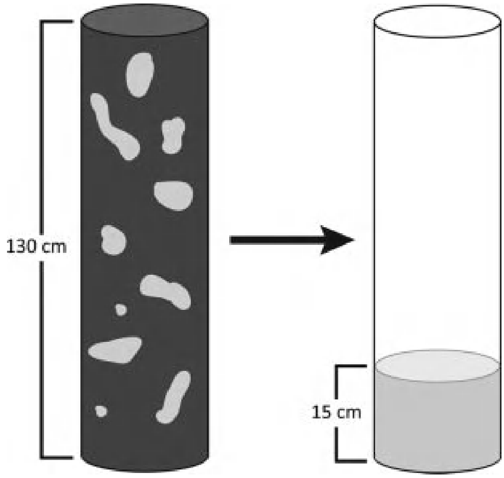

FIGURE 2.6. Soil water content in a column of soil can be expressed as an equivalent depth of water; in this case, the water content in the soil is equivalent to 15-cm depth

precipitation, streamflow, *ET*, and the change in soil moisture as depths of water (cm) over the watershed area.

Applying the Water Budget Approach

Application of a water budget as a hydrologic tool is relatively simple. If all but one component of a system can be either measured or estimated, then the unknown component can be solved directly. Because of its importance in determining the annual water yield of watersheds, *ET* often is estimated with a water budget.

A simplified water budget can be used to estimate annual ET of a watershed if changes in storage over a 1-year period are normally small. Computations for the water budget could be made beginning and ending with wet months (A–A′) or dry months (B–B′) as illustrated in Figure 2.7. The difference in soil water storage between the beginning and ending of the period should be small in either case.

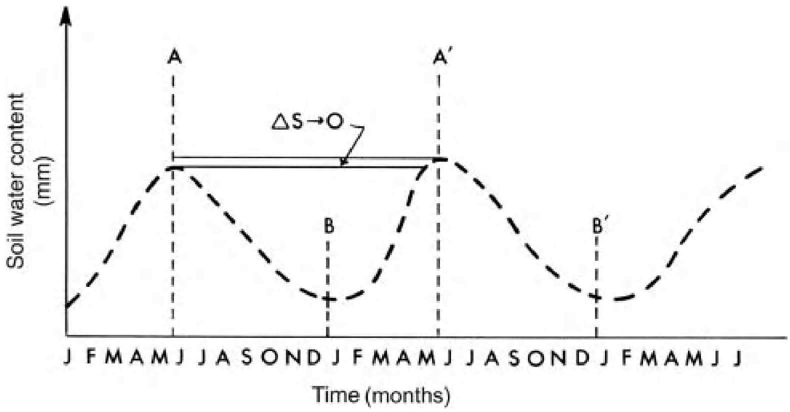

FIGURE 2.7. Hypothetical fluctuation of soil moisture on an annual basis

The above assumes that

1. All outflow of liquid water (Q) from the watershed has been measured.
2. There was no loss of water by deep seepage to underground strata ($GW_o = 0$).
3. All groundwater flow from the watershed was measured at the gauging site.
4. There is no groundwater flow into the watershed ($GW_i = 0$); therefore, $ET = P - Q$.

If geologic strata such as limestone underlie a watershed, the surface watershed boundaries might not coincide with boundaries governing groundwater flow. In such cases, there are two unknowns in the water budget, ET and groundwater seepage (GW_o), which result in:

$$ET + GW_o = P - Q \tag{2.5}$$

If groundwater seepage is suspected, it can sometimes be estimated by specialists in hydrogeology with knowledge of geologic strata and their water-conducting properties.

When annual changes in storage are considered significant, they must be determined. Estimates of change in storage become more difficult as the computational interval diminishes and as the size of the area increases. The change of storage for a small-vegetated plot can involve daily or weekly measurements of soil water content. As the size of the watershed area increases, storage changes of surface reservoirs, lakes, wetlands, and groundwater must also be taken into account. Reservoir elevation-storage data are needed to evaluate changes in lake or reservoir storage. These data are easier to analyze than storage changes in surface soils and geologic strata.

The water budget is a fundamental concept of hydrology and a useful method for the study of the hydrologic cycle. This basic equation for a water budget and its modifications allow the hydrologist to trace the pathways and changes in water storage in a watershed. Fundamental to understanding the hydrologic cycle and water budgets is an understanding of the flow of energy associated with the hydrologic cycle and the corresponding energy budget.

Energy and the Hydrologic Cycle

Energy is the capability to do work. Solar radiation that is imparted to the earth's surface is the source of energy that drives the hydrologic cycle; however, the form and status of that energy changes as water flows through the hydrologic cycle. Similar to the water budget, there is no net gain or loss of energy on earth over time, but there are exchanges of energy that involve change in state of water and transfers of energy as water moves through the hydrologic cycle.

The linkage between water and energy budgets is direct. The net energy available at the earth's surface is apportioned largely in response to the presence or absence of water. Reasons for studying energy flows and the energy budget, therefore, are to develop a better understanding of the hydrologic cycle and be able to quantify or estimate energy-dependent processes of change of state of water (evaporation, condensation, and snowmelt) and processes involving the flow of water. Although the earth's surface neither gains nor loses significant quantities of energy over long periods of time, there can be a net gain or loss for any given time interval. The following discussion emphasizes the general concepts of the energy budget.

Radiation

All substances with a temperature above absolute zero (0 K) emit electromagnetic radiation, as determined by

$$W = \varepsilon\sigma T^4 \tag{2.6}$$

where W is the emission rate of radiation in cal/cm^2/min (langleys/min); ε is the emissivity, the radiation emitted from a substance divided by the radiation emitted from a perfect blackbody (solid terrestrial objects have emissivities of 0.95–0.98, often approximated as 1.0); σ is the Stefan–Boltzmann constant (8.132×10^{-11} cal/cm^2/K^4/min); and T is the absolute temperature in K.

The amount of radiation at a particular wavelength is temperature dependent as is the emission rate of radiation. A perfect blackbody absorbs and emits radiation in all wavelengths. By convention, radiation is separated into:

1. *Shortwave* or solar radiation, sometimes called *insolation,* which includes wavelengths up to 4.0 μm.
2. *Longwave* or terrestrial, radiation, which is radiation above 4.0 μm.

As temperature increases, the greatest magnitude of emitted radiation occurs at shorter wavelengths, that is, the hotter the substance, the shorter the wavelength. Therefore, the sun emits maximum radiation at shorter wavelengths than do terrestrial objects. A doubling of the absolute temperature increases the emission rate of radiation 16-fold. The sun has a temperature of about 6000 K and emits about 10^5 cal/cm^2/min while a soil surface with a temperature of 300 K (27°C) emits about 0.66 cal/cm^2/min. The radiant environment of soil, plant, water, and snow surfaces is determined by both shortwave and longwave processes of radiation.

Shortwave radiation received at the earth's surface is composed of direct solar radiation and diffuse solar radiation. The latter represents solar radiation that is scattered and reflected by air molecules, clouds, dust, and other atmospheric particles. Diffuse skylight averages about 15% of the total downward stream of solar radiation. Any measurement of solar radiation at the earth's surface would not differentiate between direct and diffuse solar radiation; therefore, total incoming solar radiation will be referred to as W_i.

The amount of W_i that is absorbed by a terrestrial surface depends on the albedo or shortwave reflectivity of that surface. *Albedo* (α) is the proportion of the total shortwave radiation (W_i) that is reflected by an object. Light-colored surfaces have a higher albedo than dark-colored surfaces (Table 2.2). The net shortwave radiation at a surface is then determined as $(W_i) - \alpha\,(W_i)$ or $(1 - \alpha)\,(W_i)$.

The atmosphere and all terrestrial objects emit longwave radiation. The primary longwave-emitting constituents in the atmosphere are CO_2, O_3, and liquid and vapor forms of H_2O. Soil and plant surfaces reflect only a small portion of the total downward longwave radiation (W_a). Therefore, terrestrial objects are usually considered blackbodies in terms of longwave radiation. The net longwave radiation at a surface is the difference between incoming (W_a) and emitted (W_g) longwave radiation ($W_a - W_g$).

Net radiation or the net all-wave radiation (R_n) is the resulting radiant energy available at a surface:

$$R_n = (W_i)(1 - \alpha) + W_a - W_g \tag{2.7}$$

TABLE 2.2. Albedos of natural terrestrial surfaces

Terrestrial surface	Albedo (α) (%)
Fresh, new snow	80–95
Old snow	40
Dry, light sand	35–60
Dry grass	20–32
Cereal crops	25
Eucalyptus	20
Mixed hardwood forests (in leaf)	18
Rain forests	15
Pine forests	10–14
Bare wet soil	11

Source: From Lee (1980), Reifsnyder and Lull (1965), U.S. Army Corps of Engineers (USACE) (1956).

Energy Budget

Net radiation can be either positive or negative for a particular interval of time. A positive R_n represents excess radiant energy for some time interval. According to the conservation of energy principle, any excess R_n must be converted into other nonradiant forms of energy. When positive, R_n can be allocated at a surface as follows (for a snow-free condition):

$$R_n = (L)(E) + H + G + P_s \tag{2.8}$$

where L is the latent heat of vaporization (cal/g); E is the evaporation (g/cm^2 or cm^3/cm^2); H is the energy flux that heats the air or sensible heat (cal/cm^2); G is the heat of conduction to ground or rate of energy storage in terrestrial system (cal/cm^2); and P_S is the energy of photosynthesis (cal/cm^2). The latent heat of vaporization (L) and evaporation (E) is usually expressed as a product (LE), which represents the energy available for evaporating water.

Net radiation is the primary source of energy for evaporation, transpiration, and snowmelt. The allocation of net radiation in snow-free systems is dependent largely on the presence of liquid water. If water is abundant and readily available at the evaporating surface, as much as 80–90% of the net radiation can be consumed by evaporation (LE). Little energy is left to heat the air (H) or ground (G). If water is limited, a greater amount of net radiation energy is available to heat the air, the ground surface, and other terrestrial objects. Losses (or gains) of energy to the interior earth do not change rapidly with time and are usually small in relation to the net radiation. Similarly, the energy consumed in photosynthesis (of immeasurable importance to life on earth) is a small portion of the net radiation and is usually not considered in hydrologic analyses. When snow is present, the majority of net radiation can be apportioned to snowmelt (see Chapter 3).

Energy budget applications to watersheds are concerned mainly with net radiation (R_n), latent heat (LE), and sensible heat (H). As indicated in Equation 2.9, the R_n can be allocated into LE and H; however, additional energy for evaporation can be provided from sensible heat that is transferred laterally from adjacent areas, a process called *advection*. The best example of advection contributions is that of an oasis, where a well-watered plant community can receive large amounts of sensible heat from the surrounding dry and hot desert. Energy budgets of wet and dry soils are compared with an oasis in Table 2.3. Note that for the dry bare soil condition, much of the net radiation is used to heat the air

TABLE 2.3. Energy budget measurements at Akron, CO, USA, for three different conditions

	(cal/cm^2/day)			
Site description	R_n	G	H	LE
Dry bare soil	284	18	220	46
Wet bare soil	226	58	−52	220
Oasis condition	388	18	−180	550

Source: Adapted from Hanks et al. (1968), as reported by Hanks and Ashcroft (1980).
R_n, net radiation; G, heat of conduction to ground; H, sensible heat; LE, energy used in evaporation.

($H = 220$ cal/cm^2/day). In contrast, for the oasis, 180 cal/cm^2/day of H is added by advection to the oasis in addition to net radiation. Energy budget calculations for an oasis condition are presented in Box 2.2.

The height and surface area of vegetation affects the amount of energy that is available for LE. Greater amounts of energy can be received by an island of tall forest vegetation with its greater surface area than shorter herbaceous vegetation such as agricultural crops. The laterally exposed surface area of a tall forest intercepts more solar radiation and more sensible heat from moving air masses if the vegetation is cooler than the air. The total latent heat flux then is determined by

$$LE = R_n + H \quad \textbf{(2.9)}$$

Box 2.2

Energy Budget Components for Oasis Conditions at Aspendale, Australia (Adapted from Penman et al., 1967)

LE = energy used in evaporation
R_n = net radiation = 433 cal/cm^2/day
G = heat of conduction to ground = 21 cal/cm^2/day
H = sensible heat = −183 cal/cm^2/day

The energy consumed in evapotranspiration was

$$LE = R_n - G - H = 433 - 21 - (-183) = 595 \text{ cal/cm}^2\text{/day}$$

Assuming the latent heat of vaporization to be 585 cal/cm^3, the actual evapotranspiration (ET) would be

$$E = \frac{595\,\text{cal/cm}^2\text{/day}}{585\,\text{cal/cm}^3} = 1.02\,\text{cm/day}$$

Box 2.3

Energy and Water Budgets in a Changing Climate

Global warming or cooling can change the allocation of water that is stored or in circulation. Given that the earth's surface is more than 70% water, increased warming would increase evaporation rates and, to a limited extent, increase the capacity of the atmosphere to hold water in-between precipitation events. Water vapor stored in the atmosphere, however, is held temporarily, so increased evaporation must be followed by increased precipitation falling globally. The melting of ice during global warming reduces the amount of water stored in glaciers and in polar regions, but would increase the surface area of liquid water on earth which in turn can lead to greater evaporation and precipitation cycles. With greater increases in cloud cover globally, the earth's albedo would increase, thereby diminishing the amount of solar radiation reaching the earth's surface. Subsequently, global cooling could take place and lead to an expansion of polar ice, further increasing the earth's albedo and cooling.

There are many theories of why climate changes have occurred in the past (Sellers, 1965); however, with energy and water cycles on earth being dynamic and interdependent, it is clear that climate changes are inevitable. Over geologic time frames, the earth has experienced periodic ice ages and interglacial warming periods with significant changes in the type, amount, and location of water stored on the earth. In time frames of centuries, responses to the possible effects of global warming must anticipate various outcomes for which we must be able to cope.

While advection is the lateral movement of warm air to cooler plant–soil–water surfaces, *convection*, in contrast, describes the vertical component of sensible-heat transfer. The combination of advection and convection over and within irregular tree canopies can contribute significant quantities of sensible heat for evapotranspiration. Climatic variability that leads to prolonged changes in air temperatures globally can affect water budgets globally (Box 2.3).

Energy and the Flow of Liquid Water

Once liquid water reaches the earth's surface, the energy status of water is affected by the conditions present in the soil–plant system of a watershed. In discussing the energy status of water, we differentiate between energy potential and kinetic energy that can be exerted on soils, stream channels, and so forth. For example, water droplets that form in the atmosphere have potential energy based on their mass and elevation. This potential energy becomes kinetic energy (the product of mass and raindrop velocity squared) when raindrops fall to earth and impart energy to the soil surface expressed as momentum. It is these forces that cause work to be done in both the dislodgement and transport of soil particles that cause soil erosion (see Chapter 8).

After rainfall reaches the soil surface, its energy status is affected by the soil conditions; for example, if rainfall infiltrates into a dry soil, the adhesive forces in the soil attract water to the soil particles. The energy status of this soil water is reduced; that is, its ability to perform work is restricted because of the strong adhesive forces. As a result, there is capillary flow from wetter to drier areas in the soil. As the soil moisture content increases, the voids in soil become filled with water and cohesion of water molecules becomes stronger than the adhesive attraction to soil particles and, as a result, water eventually is pulled downward under the influence of gravity. If the downward movement of water becomes restricted by an impervious layer, water can become impounded above the layer; as the depth of water increases, a positive pressure of water develops. This water has the potential to perform work that exceeds that of water freely draining from the soil. To explain the processes of water flow and energy relationships in soils, we introduce the concept of water potential below.

WATER FLOW IN SOIL

Water flows through unsaturated soils in response to the difference in water potential (ψ) between two points. The water potential concept is derived from the second law of thermodynamics and relates to the free energy of water. Water always flows from a region of higher free energy (higher ψ) to a region of lower free energy (lower ψ). *Water potential* is the amount of work that a unit volume of water is capable of doing in reference to an equal unit of pure, free water at the same location in space. It can also be the minimum work needed to move a unit of water from the soil that is in excess of the work needed to move an equal unit of pure, free water from the same location in space.

A system that physically or chemically restricts the free energy of water results in negative values of water potential ($-\psi$). Gradients of negative water potentials are most common in soil–plant–atmosphere systems. The status of soil water potential (ψ_s) is determined by soil conditions that affect the free energy of soil water, which include one or more of the following potentials:

$$\psi_s = \psi_m + \psi_g + \psi_p + \psi_o + \psi_t \qquad \textbf{(2.10)}$$

where ψ_m is the matric potential; ψ_g is the gravitational potential; ψ_p is the pressure potential; ψ_o is the solute, or osmotic, potential; and ψ_t is the thermal potential.

Water content in the soil determines which of the three potentials, ψ_m, ψ_g, or ψ_p, controls the flow of water in a soil (Fig. 2.8). When a soil becomes saturated and water becomes ponded above the soil surface or within the soil, the pressure (P) exerted by the column of water above some point in the soil is directly proportional to the height of the water column (h_p) and can be approximated by

$$P = \rho g h_p \qquad \textbf{(2.11)}$$

where ρ is the density of water and g is the force of gravity.

The relationship in Equation 2.11 explains why a swimmer who dives into the water feels increasing pressure with increasing depth. For a point in a saturated soil or in groundwater, the energy status of the water is characterized as pressure potential (ψ_p), which is positive; that is, it has the potential to perform work. In the example in the top frame of

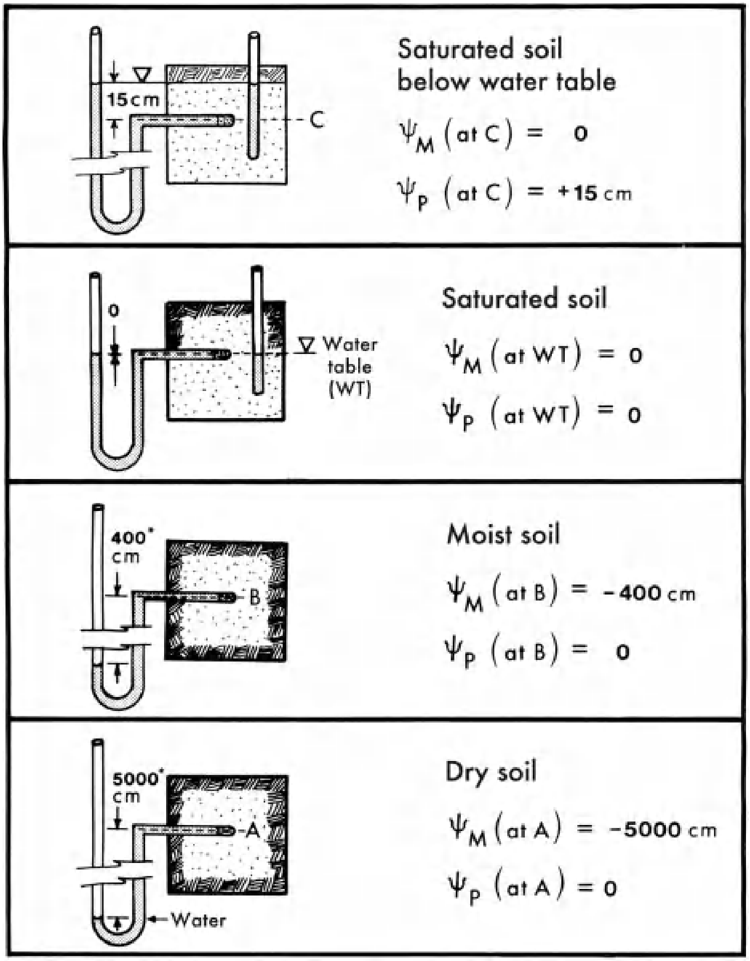

FIGURE 2.8. Matric potential (ψ_m) and pressure potential (ψ_p) of a dry soil (A), moist soil (B), at the water table (WT), and in a saturated soil (C) measured with a water manometer or tensiometer (adapted from Hanks and Ashcroft, 1980). *Vertical distance not to scale; porous ceramic coup at A, B, WT, and C

Figure 2.8, the pressure potential (ψ_p at *C*) corresponds to the depth below the free-water surface, a depth of 15 cm. If the water potential was measured at the surface of the saturated layer of soil or at the water table of a shallow groundwater aquifer, it would have no positive pressure potential and $\psi_p = 0$ as shown in the second frame in Figure 2.8.

If there is no impeding layer to prevent the water from draining out of a saturated soil, water will drain under the influence of gravity. The gravitational potential exerts a downward pressure as a function of the weight of water as determined by the height of the water column, gravity, and density. The ψ_g is the difference in elevation of a point in the system with respect to a stable reference datum, often mean sea level. When soil water content exceeds the FC of the soil, the adhesive forces that attract water to soil particles become less important in governing water flow with gravity dominating water flow through the soil, that is, it drains. The energy status of water flowing through a saturated flow would equal the sum of gravitational and pressure potential, called the *total hydraulic potential*.

When the soil moisture content is at or below FC, the water retained in the soil is due to the physical attraction of water to soil particles by both capillary and adsorptive forces that exceed the force of gravity. As a result, the free energy status of soil water becomes negative and is governed by the soil matric potential (ψ_m). As the soil dries, the matric forces in the soil increase (the magnitude of negative ψ_m increases) and more energy must be exerted to move a quantity of water in a drier soil than in a more moist soil. Because water always moves from a higher to lower energy state, water moves from a smaller negative ψ_m to a larger negative ψ_m. For example, if the moist and dry soils in the lower two frames of Figure 2.8 were connected, water would flow from soil B to soil A.

Units Commonly Used to Express Water Potential

$$\begin{aligned} 1 \text{ kilopascal (kPa)} &= 10 \text{ millibars (mb)} \\ &= 10.2 \text{ cm } H_2O \\ &= 0.01 \text{ atmosphere (atm)} \\ &= 75 \text{ cm Hg} \end{aligned}$$

Solutes in soil water create an osmotic potential (ψ_o). Solutes lower soil water potential (ψ_s) because they attract water molecules in the form of hydrated shells. This component normally has little effect on liquid-water flow in soils except at the soil–root interface. Here, a semipermeable membrane (permeable only for water molecules) must be crossed for water to enter the root vascular system.

Thermal potential (ψ_t) is usually neglected. It would appear, however, that soil temperature gradients (over time and space) would affect ψ_s under certain conditions. Higher thermal energy would increase ψ_s and, thereby, change the flow of water in soil.

Water flow in unsaturated soils is primarily a function of matric potential gradients ($d\psi_m/dx$), which are directly related to gradients in soil water content. As the roots take up water, the soil water content immediately adjacent to the roots is depleted. The lower moisture content results in a greater attraction between water and the soil particles next to the root. A gradient in soil water content and, hence, water potential is then established. Water flow is from a region of higher ψ_s to a region of lower ψ_s in all cases. The driving forces operating in unsaturated flow have been described, but the velocity of soil water flow (v) is determined by

$$v = \frac{\Delta\psi}{r_s} = -k_v\left(\frac{d\psi}{dx}\right) \quad \textbf{(2.12)}$$

where v is the velocity (cm/sec); $\Delta\psi$ is the water potential difference (kPa); r_s is the resistance of any component (kPa $\times$ sec/cm); k_v is the hydraulic conductivity (cm/kPa $\times$ sec); x is the distance over which gradient is present (cm); and $d\psi/dx$ is the total water potential gradient.

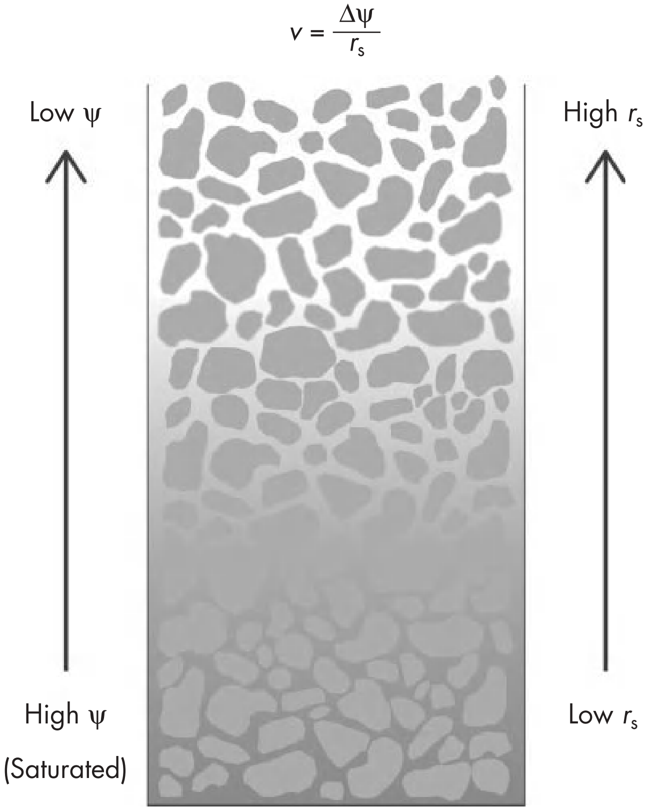

FIGURE 2.9. Water flow through soil moves from a higher water potential (ψ) or wetter soil to a lower water potential or drier soil. As the soil dries at the soil surface (or at the soil–root interface), the resistance to water flow (r_s) increases as a result of the more tortuous pathway of flow through the thinner water films surrounding soil particles

The relationship shown in Equation 2.12 is analogous to Ohm's law for electrical current. The velocity of water flow is proportional to $d\psi/dx$ and k_v. The pore size, pore geometry of the soil, and the soil water content affect the value of k_v for unsaturated conditions; therefore, k_v is not a constant in unsaturated soils. Soil texture and structure affect soil water flow in unsaturated and saturated conditions such as groundwater flow.

Water always flows from higher potential to lower potential (wetter zones to drier zones) in the soil (Fig. 2.9). The gradient is maintained in part by evaporation processes (see Chapter 4) at the soil surface or by plant transpiration in which soil moisture is taken up by plant roots and the soil dries at the soil–root interface. As the soil dries, the films around soil particles become thin and the pathway through which water flows becomes more tortuous, which is represented by an increasing resistance to flow (r_s). Water flow through soils is relatively passive until intercepted by the roots of plants. Once the root absorbs soil water, different forces become operative as the major constituents of water potential in the plant (discussed in Chapter 4).

Water Flow on Land and in Stream Channels

Excess water on the land or in stream channels flows in response to *hydraulic potential* and energy gradients expressed as slope. If there is no slope, there is no energy gradient and water will not flow. In such cases, excess water will occur as a pool above the soil surface. As water flows into a stream channel, it is a continuous flow or a stream if the energy gradient is continuous. Therefore, a flowing stream is a response to a continuous energy gradient.

The energy gradient that drives the stream is a potential energy gradient. The potential energy of water exists by virtue of its elevation, expressed as

$$\text{Potential energy} = (\text{mass}) \times (\text{gravity}) \times (\text{elevation}) \quad \mathbf{(2.13)}$$

When rainfall creates excess water in the higher elevation headwaters of a river basin, this water has a higher potential energy than excess water located downstream in the basin. Water that is ponded at upper portions of a watershed has a high potential energy but cannot perform work until it is released. The Law of Entropy tells us that systems tend toward their lowest possible homogeneous energy state, that is, their base level. This base level is sea level in the hydrologic cycle. When there is a slope or energy gradient, the force of gravity drives the stream from a higher energy state to a lower energy state by converting the stream's potential energy to kinetic energy (K_e) by virtue of the velocity of flowing water:

$$K_e = 0.5 \times (\text{mass}) \times (\text{velocity})^2 \quad \mathbf{(2.14)}$$

As water moves toward a lower energy state, it must release energy. Energy is lost as heat from friction within the flowing water, from channel friction, and when water performs work on its channel and sediments. Streams do not need to erode channels to dissipate excess energy because they can dissipate energy internally such as by turbulent flow. The turbulence in a stream is seen as the eddies, boils, and rolls of a stream that are the result of internal transfers of momentum between adjacent parcels of water.

When a stream loses energy to its channel, it does so by transferring momentum from itself to its channel. A change in momentum is a force (F) where

$$F = \Delta\,\text{momentum} = \Delta\,(\text{mass}) \times (\text{velocity}) \quad \mathbf{(2.15)}$$

When a stream applies enough force to its channel, some material will be eroded and carried away, that is, it is performing *work*. Work is defined as a force acting against a resistance to produce motion in a body. Over time, a stream forms its channel by performing work. The amount of work that a stream can perform on its channel is the stream's *power*. These concepts provide the foundation for our discussions on the processes of water flow though soils and in stream channels and the relationships of sediment flow and channel forming processes later in this book.

Summary and Learning Points

The hydrologic cycle represents the circulation of water on earth and is driven ultimately by solar radiation. Hydrologic processes involved in the hydrologic cycle involve changes in state of water and flows of water that are a function of the properties of water and the allocation of energy. Water circulates continuously over time; thus, the molecules of water that you drink today may have at one time been stored in a snowpack in the Himalayan Mountains, or fallen as rain on Inca civilizations in the Andes Mountains centuries ago. Understanding the hydrologic cycle and its linkages to energy provide the foundation for the study of hydrology.

After reading this chapter, you should be able to

1. Define the key processes in the hydrologic cycle.
2. Understand the concept and application of a water budget for a watershed.
3. Define and explain the components of an energy budget.
4. Know how an energy budget can be used to solve for components of the water budget.
5. Understand the properties of cohesion, adhesion, and capillarity of water and describe how they influence water potential and soil water movement.
6. Understand and explain the energy conditions that affect the flow of excess water in a watershed and stream channel.

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CHAPTER 3 Precipitation

INTRODUCTION

Precipitation affects the amount, timing, spatial distribution, and quality of water added to a watershed from the atmosphere. Hydrologists view precipitation as the major input to a watershed and a key to its water-yielding characteristics. Ecologists recognize the role of precipitation in determining the types of soils and vegetation that occur on a watershed. Farmers, foresters, and rangeland managers view precipitation as an essential ingredient to vegetative production on the land. The annual amount of precipitation and the high and low extremes of precipitation affect people and their use of land and water. The uncertainty of floods and droughts becomes apparent when we examine climatic variability of the past. Climatic change and variability add to this uncertainty. Most people have only a cursory understanding of the variability of precipitation, why precipitation occurs, and why it occurs where it does. Consequently an understanding of precipitation is fundamental to the study of hydrology and integrated watershed management.

In this chapter we discuss the measurement of precipitation and analysis of data that are important to the day-to-day activities of people and for the planning of future development and management of land and water resources. Rainfall affects more people globally than snow; however, information about snowfall, snow accumulation, and melt is needed on watersheds that are located in higher latitudes and at high elevations of mountainous regions. The process of precipitation, its occurrence in time and space, and the methods of measuring and analyzing rainfall and snowmelt inputs to watersheds are discussed in this chapter.

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

PRECIPITATION PROCESS

Precipitation is the result of meteorological factors and is, therefore, largely outside human control. Air masses take on the temperature and moisture characteristics of underlying surfaces, particularly when they are stationary or move slowly over large water or land surfaces. Air masses moving from an ocean to land bring to that land surface a source of moisture. Air masses from polar regions will be dry and cold. Air masses from the humid tropics are warm and moist. The pathways of jet streams in the upper atmosphere govern the movement of air masses and thus affect the amount and distribution of precipitation. Movement of air masses modifies the temperature and moisture characteristics of the atmosphere over a watershed and determines the climatic and, more specifically, the precipitation conditions that occur. Precipitation occurs in many forms (see Box 3.1) but most commonly occurs as either rain or snow, depending on air temperature. The relationship between atmospheric moisture and temperature must be understood to understand the precipitation process.

The relationships among moisture content in the atmosphere, temperature, and vapor pressure determine the occurrence and amounts of evaporation and precipitation (Fig. 3.1). A parcel of unsaturated air (point A in Fig. 3.1) can become saturated by either cooling (A to C) or the addition of moisture to the air mass (A to B). Unsaturated air can be characterized by its *relative humidity* – the ratio of vapor pressure of the air to saturation vapor pressure at a specified temperature, which is expressed as a percentage. For example, at 25°C in Figure 3.1, the relative humidity of air parcel A is 100(17 mb/31 mb) = 55%. The difference between the saturation vapor pressure (e_s) and the actual vapor pressure of

Box 3.1

Forms of Precipitation

- Drizzle – waterdrops less than 0.5 mm in diameter; intensity less than 1 mm/h
- Rain – waterdrops greater than 0.5 mm in diameter; upper limit is 6 mm in diameter
- Sleet – small frozen raindrops
- Snow – ice crystals formed in the atmosphere by the process of sublimation
- Hail – ice particles greater than 0.5 mm in diameter formed by alternate freezing and thawing in turbulent air currents; usually associated with intense convective cells
- Fog, dew, and frost – not actually precipitation; the result of interception, condensation, or sublimation; can be important sources of moisture to watersheds in coastal areas and other areas subjected to persistent fog and/or clouds

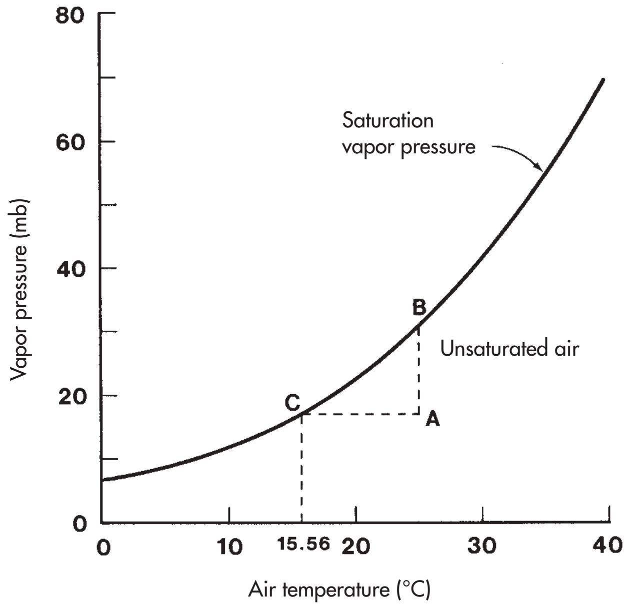


FIGURE 3.1. The relationship of saturation vapor pressure and air temperature. A parcel of unsaturated air **(A)** must be cooled **(A to C)** or moisture must be added **(A to B)** before saturation occurs

an unsaturated air parcel (e_a) is the *vapor pressure deficit* (for A it is 31 mb − 17 mb = 14 mb). The temperature at which a parcel of unsaturated air reaches its saturation vapor pressure is the *dew point temperature*, 15.56°C in Figure 3.1 for the air parcel A.

Precipitation occurs when the following three conditions are met:

(1) The atmosphere becomes saturated.
(2) Small particles or nuclei such as dust or ocean salt are present in the atmosphere upon which condensation or sublimation can take place.
(3) Water or ice particles coalesce and grow large enough to reach the earth against updrafts.

Saturation results when either the air mass is cooled until the saturation vapor pressure is reached or moisture is added to the air mass (Fig. 3.1). Rarely does a direct introduction of moist air cause precipitation. More commonly, precipitation occurs when an air mass is lifted in elevation, becomes cooled, and reaches its saturation vapor pressure. Air masses are lifted as a result of frontal systems, orographic effects, and convection. Different storm and precipitation characteristics result from each of these lifting processes (Fig. 3.2).

Frontal precipitation occurs when general circulation brings two air masses of different temperature and moisture content together and air becomes lifted at the frontal surface. A cold front results from a cold air mass replacing and lifting a warm air mass, which already has a tendency to rise. Cold fronts are characterized by high-intensity rainfall of relatively short duration and usually cover an area narrower than that of warm fronts. Conversely,

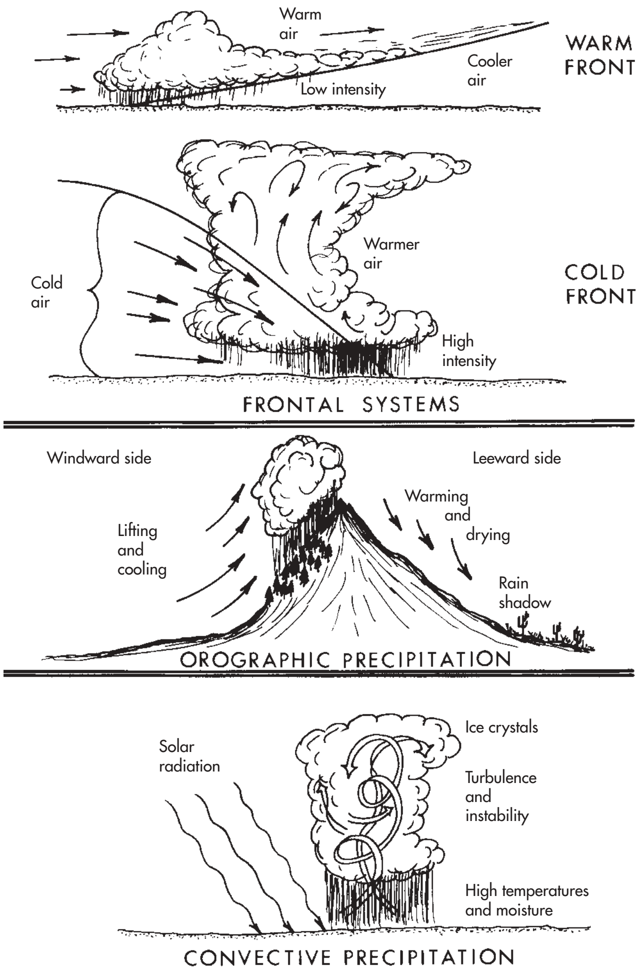


FIGURE 3.2. Mechanisms causing air masses to lift in elevation and cool and result in precipitation

a warm front results when warm air rides up and over a cold air mass. A warm front is generally characterized by widespread, low-intensity rainfall.

Orographic precipitation occurs when general circulation forces an air mass up and over a mountain range. As the air mass becomes lifted, a greater volume of the air mass reaches saturation vapor pressure, resulting in more precipitation with increasing elevation. A striking example of orographic precipitation occurs on the windward side of Mount Rainier, Washington (USA), where annual precipitation varies from about 1345 mm at 305 m elevation to 2921 mm at 1690 m. Once an air mass passes over a mountain, a lowering and warming of the air occurs, creating a dry, rain shadow effect on the leeward side. Annual precipitation averages less than 173 mm in the rain shadow area near Priest Rapids Dam in central Washington.

Convective precipitation, characterized by summer thunderstorms, is the result of excessive heating of the earth's surface. When the moist air immediately above the surface becomes warmer than the air mass above it, lifting occurs. As the air mass rises, it cools and condensation takes place, releasing the latent heat of vaporization which adds more energy to the rising air mass, and, consequently, more lifting occurs. Rapidly uplifted air can reach high altitudes where water droplets become frozen and hail forms or becomes intermixed with rainfall. Such rainstorms or hailstorms are some of the most severe precipitation events anywhere and are characterized by high-intensity, short-duration rainfall over relatively limited areas. Flash flooding can result when thunderstorms occur over a large enough area.

RAINFALL

Rainfall is a meteorological event throughout most of the world. It is the primary input in the hydrologic cycle of many watersheds and larger river basins. While rainfall is often a key source of water to people, excessive rainfall amounts can lead to downstream flooding. It is necessary, therefore, that hydrologists and watershed managers know how to measure rainfall and analyze the measurements that are obtained.

Methods of Rainfall Measurement

Measurements of rainfall are needed for many hydrologic applications that are discussed in this book. Weather forecasting and flood forecasting require estimates of rainfall occurring in "real time," meaning as it falls to the earth's surface, or shortly thereafter. Radar provides qualitative estimates of rainfall amounts and intensities over large areas during a storm, provided that radar coverage is sufficient. Radar senses the backscatter of radio waves caused by water droplets and ice crystals in the atmosphere. The area of coverage and relative intensity rainfall can be estimated up to a distance of 250 km. Radar echoes can often be correlated with measured rainfall, but this calibration is hampered by ground barriers, raindrop size, distribution of rainfall, and other storm factors resulting in large errors (Sauvageot, 1994). An evaluation of errors associated with the WRS-88D Precipitation Processing Subsystem (PPS) of the National Weather Service (USA) indicates that there are potential remedies to overcome the major errors associated with radar (http://www.srh.noaa.gov/mrx/research/precip/precip.php). Readers are encouraged to search the National Weather Service website for applications of radar to estimate rainfall (see www.nws.noaa.gov/).

Hydrologic studies of rainfall–streamflow relationships, for example, post-flood analyses, and the testing and application of hydrologic models require accurate measurements of rainfall on a watershed. The most common method of measuring rainfall is to use a series of gauges – typically cylindrical containers 20.3 cm (8 in.) in diameter (Fig. 3.3a). Three types of gauges in general use are the standard gauge, the storage gauge, and the recording gauge. Standard or nonrecording gauges are often used because of economy. Such gauges should be read periodically, normally every 24 hours at the same time each day. The standard gauge magnifies rainfall depth 10-fold because it funnels rainfall into an internal cylinder of 10-fold smaller cross-sectional area. Storage gauges have the same size opening as standard gauges but have a greater storage capacity, usually 1525–2540 mm of

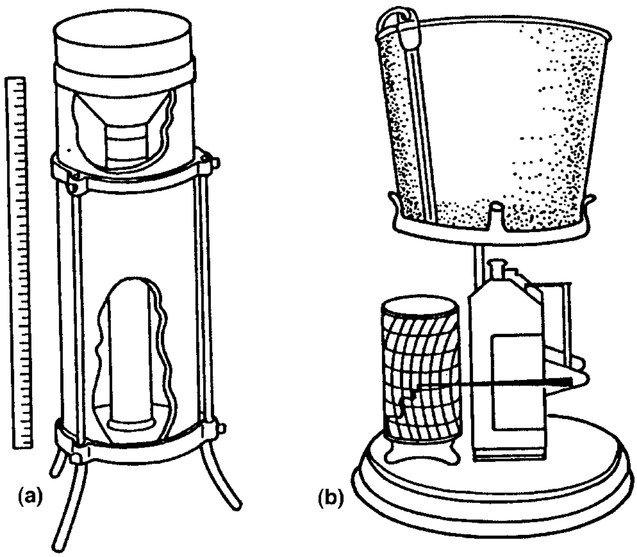


FIGURE 3.3. Types of rain gauges. **(a)** Cutaway of the standard (Weather Service) gauge. **(b)** A weighing-type recording gauge with its cover removed to show the spring housing, recording pen, and storage bucket (from Hewlett, 1982, © University of Georgia Press, with permission)

rainfall. These gauges can be read periodically, for example, once a week, once a month, or seasonally. A small amount of oil is usually added to gauges that are read less frequently than every 24 hours to suppress evaporation.

The use of recording gauges, which allow for continuous measurement of rainfall, is more limited because of their higher cost. Examples of recording rain gauges are the weighing-type (Fig. 3.3b) and the tipping-bucket gauge (Fig. 3.4). The weighing-type gauge records the weight of water with respect to time by means of a calibrated pen on a clock-driven drum. The chart on the drum indicates the accumulated rainfall with time. Rainfall intensity is obtained by determining incremental increases in the amount per unit of time (typically 1 hour). A tipping-bucket gauge records intensity, making a recording each time a small cup (usually 1 mm capacity) fills with water and then empties as it tips back and forth. Since about 0.2 seconds is required for the bucket to tip, high-intensity rainfall events might not be accurately measured.

The accuracy of rainfall measurements is affected by both gauge site characteristics and the relationship of the location of gauges to the watershed area. As a general rule, a rain gauge should be located in a relatively flat area with the funnel opening in a horizontal plane. The standard procedure in the United States is to situate the gauge so that the funnel orifice is 1 m above the ground surface. The gauge should also be far enough away from surrounding objects so that the rainfall catch is not affected by the objects. A clearing defined by a 30–45° angle from the top of the gauge to the closest object is usually sufficient (Fig. 3.5). If the gauge is located too close to trees or structures, the wind patterns circulating around the gauge can result in catches far different from the rainfall that actually occurred. Ideally, one should select small openings in a forest or other areas that are sheltered from

FIGURE 3.4. Tipping-bucket gauge

the full force of the wind but that meet the above criteria. A gauge might be located in a large enough opening when it is initially installed but then it can become affected by forest growth in adjacent areas over time. This possibility is particularly important in areas where vegetation grows quickly. High wind speeds also diminish the efficiency of a gauge catch (Table 3.1). Wind shields, such as Nipher or Alter shields, should be used to reduce eddy effects in areas of high wind speeds. Errors caused by catch deficiencies are called *instrumentation errors*.

Gauges should be located throughout watersheds so that spatial and elevational differences in rainfall can be measured. Such factors as topographic barriers, elevational differences, and storm-track patterns or tracking should be considered in developing a rain gauge network. Practical considerations including economics and accessibility usually limit the type, number, and location of gauges.

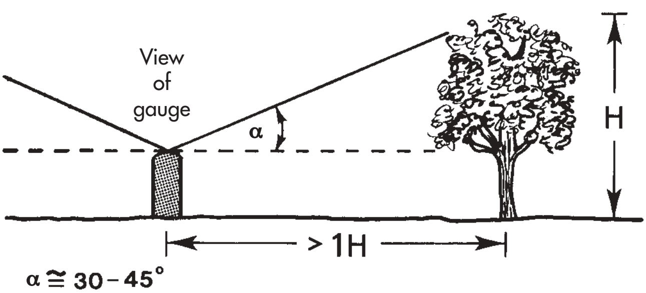


FIGURE 3.5. Proper siting of a rain gauge with respect to the nearest object

Table 3.1. Effect of wind velocities on the catch of precipitation by standard rainfall gauges

	Catch deficiency (% of true amount)	
Wind velocity (mile/h)	Rainfall	Snowfall
0	0	0
5	6	20
15	26	47
25	41	60
50	50	73

Source: From Gray (1973), with permission.

Number of Gauges Required

The number of gauges required to measure rainfall should generally increase with the size of the watershed and with the variability of precipitation. Sampling requirements can be determined by standard statistical methods (Haan, 2002). Use of random sampling as a means of excluding bias in the selection of gauge sites and estimating the number of gauges needed is often suggested. Random sampling of rainfall might not be practical on most watersheds because the presence of dense forest cover, steep topography of the watershed, and so forth, can limit site locations for rain gauges. Accessibility can limit the preferred siting of gauges in remote watersheds or where access to private property is denied. Insufficient sampling is more often the norm than the exception as a consequence of these conditions.

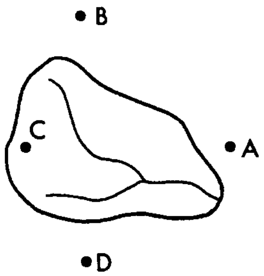
Using measurements of rainfall from a network of standard gauges that are read after each storm event can provide an estimate of rainfall variability on a watershed for monthly, seasonal, or annual periods. The effects of different storm types on this variability can be lost by reading storage gauges monthly or seasonally, but systematic differences in precipitation between parts of the watershed for these longer periods can be estimated.

Calculating Mean Rainfall on a Watershed

The mean depth of rainfall on a watershed is required in many hydrologic investigations. Several methods are used in deriving this value. The three most common are the arithmetic mean, the Thiessen polygon, and the isohyetal methods (Fig. 3.6).

Arithmetic Mean

A straightforward calculation of the arithmetic average is the simplest of all methods for estimating the mean rainfall on a watershed (Fig. 3.6). This method provides good estimates if the gauges are numerous and uniformly distributed. Even in mountainous regions, averaging the catch of a dense rain gauge network will yield good estimates if the orographic influence on rainfall is considered in the selection of gauge sites. However, if gauges are relatively few, irregularly spaced, or rainfall varies considerably over the watershed, more sophisticated methods are warranted.



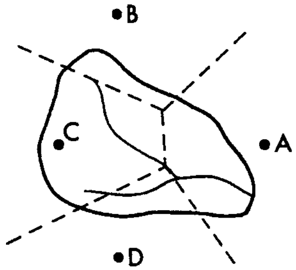
Arithmetic Mean Method

Source Data: Daily rainfall measured at each gauge in centimeters

A	B	C	D
4	8	10	6

$$\text{Arithmetic mean} = \frac{4+8+10+6}{4}$$

$$= 7\text{ cm}$$

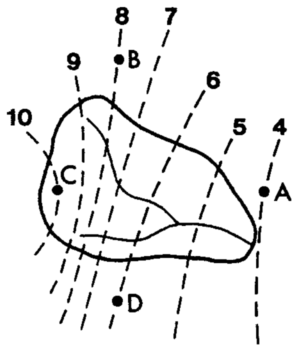


Thiessen Polygon Method

Thiessen Polygon:

Station	Depth (cm)		Area in Polygon*		Volume (cm)
A	4	×	0.28	=	1.12
B	8	×	0.09	=	0.72
C	10	×	0.49	=	4.90
D	6	×	0.14	=	0.84
			Sum	=	7.58

*As a fraction of total area.



Isohyetal Method

Isohyetal:

Mean Depth (cm)		Area between Isohyets*		Volume (cm)
4.5	×	0.12	=	0.54
5.5	×	0.25	=	1.38
6.5	×	0.14	=	0.91
7.5	×	0.13	=	0.98
8.5	×	0.18	=	1.53
9.5	×	0.14	=	1.33
10.5	×	0.04	=	0.42
		Sum		7.09

*As a fraction of total area.

FIGURE 3.6. Arithmetic mean, Thiessen polygon, and isohyetal methods of calculating the mean rainfall on a watershed

Thiessen Polygon Method

When gauges are nonuniformly distributed over a watershed, the Thiessen polygon method can improve estimates of rainfall amounts over the entire area. Polygons are formed from the perpendicular bisectors of lines joining nearby gauges (Fig. 3.6). The watershed area within each polygon is determined and is used to apportion the rainfall amount of the gauge in the center of the polygon. It is assumed that the depth of water recorded by the rain gauge located within the polygon represents the depth of rain over the entire area of the polygon. The results obtained are usually more accurate than the arithmetic average when the number of gauges on a watershed are limited and when one or more gauges are located outside the watershed boundary.

The Thiessen method allows for nonuniform distribution of gauges but assumes linear variation of rainfall between gauges and makes no attempt to allow for orographic influences. Once the area-weighing coefficients are determined for each station, they become fixed and the method is as simple to apply as the arithmetic method.

Isohyetal Method

With the isohyetal method, gauge location and amounts are plotted on a suitable map, and contours of equal rainfall (isohyets) are drawn (Fig. 3.6). Rainfall measured within and outside the watershed can be used to estimate the pattern of rainfall, and isohyets are drawn according to gauge catches. The average depth is then determined by computing the depth between isohyets on the watershed and dividing by the total area. Many hydrologists believe that this is theoretically the most accurate method of determining mean watershed precipitation, but it is also by far the most laborious.

The isohyetal method is particularly useful when investigating the influence of storm patterns on streamflow and for areas where orographic rainfall occurs. In some instances, relationships between rainfall and elevation can be used advantageously, and only a few gauges need to be set out such as at accessible elevations. Where orographic rainfall occurs, contour intervals can sometimes be used to help estimate the location of the lines of equal rainfall. Precipitation amounts are then determined for each elevation zone or band, and the respective areas are weighed to obtain estimates for the entire watershed. Rainfall–elevation relationships would be different for watersheds on the windward side of a mountain range than on the leeward side of a mountain range.

The accuracy of the isohyetal method depends largely on the skill of the analyst. An improper analysis can lead to serious errors. The results obtained with the isohyetal method will be essentially the same as those obtained with the Thiessen method if a linear interpolation between stations is used.

Other Methods

Other methods of calculating the mean rainfall on a watershed and examples of their application can be found on the website for the National Weather Service (NWS) River Forecast Center (http://srh.noaa.gov/abrfc/?n=map).

Errors Associated with Rainfall Measurement

Hydrologic studies of watersheds are often constrained by an inadequate number of rainfall gauges, the absence of long-term precipitation records, or both. Two types of errors should be considered when obtaining rainfall measurements. *Instrumentation error* is related to the accuracy with which a gauge catches the true rainfall amount at a point while *sampling error* is associated with how well the gauges on a watershed represent the rainfall over the entire watershed area.

Taking care in siting a gauge correctly and proper maintenance can minimize instrumentation errors. For standard 20.3 cm (8 in.) precipitation gauges in the United States, there are biases in gauge catch from the effects of wind and from wetting of the gauge orifice. Legates and DeLiberty (1993) reported up to 18 mm per month of systematic undercatch bias for winter precipitation in the northeastern and northwestern United States. A 28% undercatch bias was reported for winter precipitation in the northern Rocky Mountains and Upper Midwest. The biases can be corrected by using

$$P_c = k_r\left(P_{gr} + dP_{wr}\right) + k_s\left(P_{gs} + dP_{ws}\right) \tag{3.1}$$

where P_c is the gauge-corrected precipitation; P_g is the measured precipitation; k is the wind correction coefficient (usually ≥ 1); dP_w is the correction for wetting loss (usually 0.15 for rain events and half this value for snowfall); and subscripts "r" and "s" denote rainfall and snowfall, respectively. The wetting loss refers to the amount of rain or snow that is initially stored on the interior surface of the rain gauge at the beginning of the storm.

The wind correction coefficient for rainfall (Sevruk and Hamon, 1984) is

$$k_r = \frac{100}{100 - 2.12w_{hp}} \tag{3.2}$$

and for snowfall (Goodison, 1978) it is

$$k_s = e^{0.1338}w_{hp} \tag{3.3}$$

where w_{hp} is the wind speed at the height of the gauge orifice.

Properly designing a network with an adequate number of gauges, on the other hand, minimizes sampling error. Accessibility and economic considerations usually determine the extent to which sampling errors can be reduced.

Analysis of Rainfall Measurements

Once we have obtained rainfall records, a number of analyses can be performed to enhance our knowledge of hydrology and climate. This section of the chapter describes some of the more common types of rainfall analysis.

Estimating Missing Data

One or more gauges in a rainfall network can become nonfunctioning for a period of time. One way to estimate the missing record for such a gauge is to use existing relationships with adjacent gauges. For example, if the rainfall for a storm is missing for station C and if rainfall at C (seasonal or annual) is correlated with that at stations A and B, the normal ratio method can estimate the storm rainfall for station C:

$$P_C = \frac{1}{2}\left(\frac{N_C}{N_A}P_A + \frac{N_C}{N_B}P_B\right) \tag{3.4}$$

where P_C is the estimated storm rainfall for station C (mm); N_A, N_B, and N_C are normal annual (or seasonal) rainfall for stations A, B, and C (mm), respectively; and P_A and P_B are storm rainfall for stations A and B (mm), respectively.

The generalized equation for missing data is

$$P_x = \frac{1}{n}\left(\frac{N_x}{N_1}P_1 + \cdots + \frac{N_x}{N_n}P_n\right) \tag{3.5}$$

Equation 3.5 is recommended only if there is a high correlation with other stations.

Double Mass Analysis

Double mass analysis is a method of checking the consistency of measurements at a rainfall station against those of one or more nearby stations. An example helps to explain the application of this method. Consider that station E has been collecting rainfall data for 45 years. Originally, the station was located in a large opening in a conifer forest, but over the years the surrounding forest has grown to the point where you suspect that the catch of this gauge is now affected. Based on one's knowledge of the precipitation patterns in the region, you recognize that the same storm patterns influence stations H and I, although their elevational differences and other factors cause annual rainfall to differ. There was a consistent correlation between the average of stations H and I and that of E in the early years of station E. By plotting the accumulated annual rainfall of E against the accumulated average annual rainfall of H and I, we find that the relationship clearly changed after 1970 in this example (Fig. 3.7). The relationship then can be used to correct existing rainfall catch at E so that it better represents the true catch at the location without the interference of nearby trees.

Frequency Analysis

Water resource systems, such as waterways, small reservoirs, irrigation networks, and drainage systems for roads, should be planned and designed for future precipitation events,

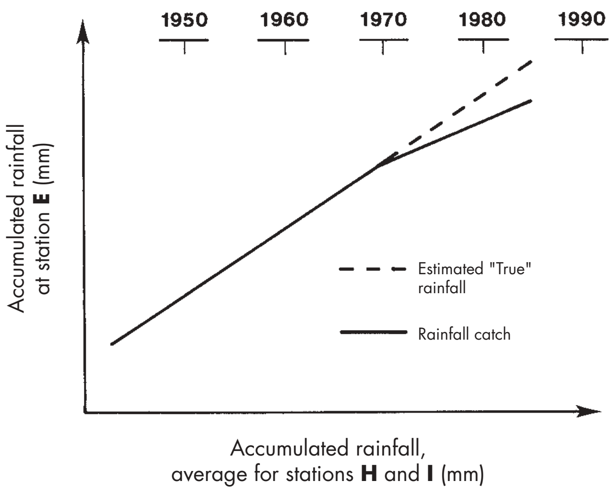


FIGURE 3.7. A double mass plot of the annual rainfall at station E relative to the average annual rainfall at stations H and I

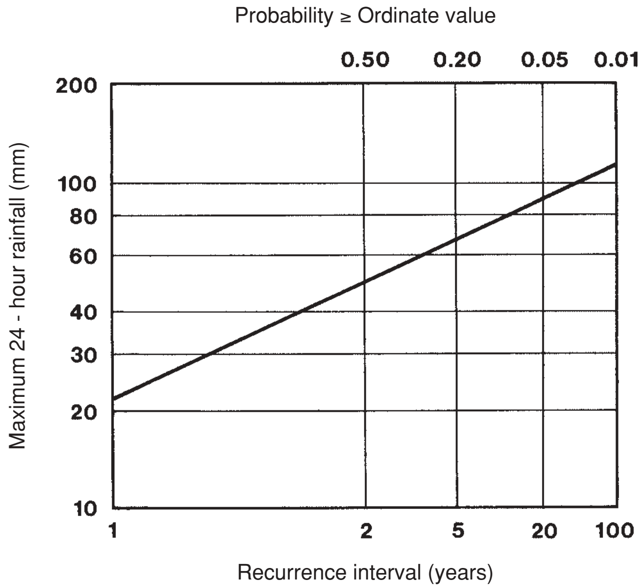


FIGURE 3.8. Frequency curve of the daily rainfall recorded for a single station

the magnitude of which cannot be accurately predicted. Weather systems vary from one year to the next, and no one can accurately predict what the next year, season, or even month will bring. Therefore, we rely on statistical analyses of rainfall amounts over certain periods. From these analyses, the frequency distributions of past events are determined and cumulative frequency curves are developed to facilitate estimating the probability or likelihood of having certain events occur or be exceeded over a specified period. This approach, called *frequency analysis*, will be discussed here for rainfall events.

The objective of frequency analysis is to develop a frequency curve, which is a relationship between the magnitude of events and their associated probabilities. Maximum annual rainfall amounts of specified durations can be tabulated, and a frequency curve can be developed for each duration. For example, the annual maximum 1-hour, 3-hour, 6-hour, 12-hour, and 24-hour duration rainfall amounts are identified (preferably from long-term rainfall records) and frequency curves developed for each duration. Figure 3.8 illustrates a frequency curve for the maximum 24-hour or daily rainfall at a particular location. Details of developing a frequency distribution function of rainfall data are outlined by Haan (2002) and other references on statistical hydrology. The recurrence interval (T_r) can be approximated by

$$T_r = \frac{n+1}{m} \tag{3.6}$$

where n is the number of years of record and m is the rank of the event.

The recurrence interval (*RI*), or return period, which corresponds to a specified probability (p), is determined as the reciprocal of the probability:

$$T_r = \frac{1}{p} \tag{3.7}$$

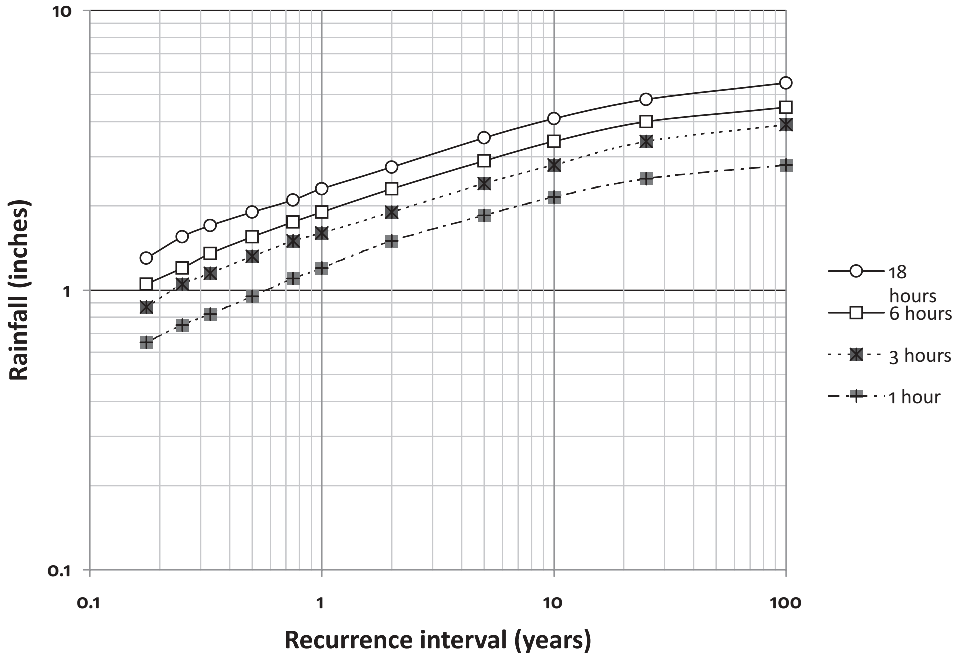


FIGURE 3.9. Intensity duration frequency curves for the Midwestern United States (from Huff and Angel, 1992)

For example, in Figure 3.8 the probability of having a 24-hour rainfall of 90 mm or more in a year is about 0.05; its associated *RI* would be 20 years. Therefore, one could say that the 20-year, 24-hour maximum rainfall is 90 mm. Regional frequency curves have been developed for different durations of rainfall amounts as depicted in Figure 3.9 for the Midwestern United States. Such rainfall-intensity–duration curves are used for various hydrologic analyses as described in Chapter 6.

In addition to developing rainfall frequency curves based on maximum annual events of different durations, they also can be developed to evaluate minimal rainfall amounts over extended periods of time to analyze the frequency of certain droughts. For example, a frequency curve can be developed for the minimum 12-month rainfall amounts.

Once a frequency curve is developed, the probability of exceeding certain rainfall amounts over some specified period can be determined. The probability that an event with probability p will be equaled or exceeded x times in N years is determined by

$$\text{Prob}\,(x) = \frac{N!}{x!(N-x)!}(p)^x(1-p)^{N-x} \tag{3.8}$$

This relationship can be simplified by considering the probability of at least one event with probability p being equaled or exceeded in N years as follows:

$$\text{Prob}\,(\text{no occurrences in } N \text{ years}) = (1-p)^N \tag{3.9}$$

Therefore,

$$\text{Prob}\,(\text{at least one occurrence in } N \text{ years}) = 1-(1-p)^N \tag{3.10}$$

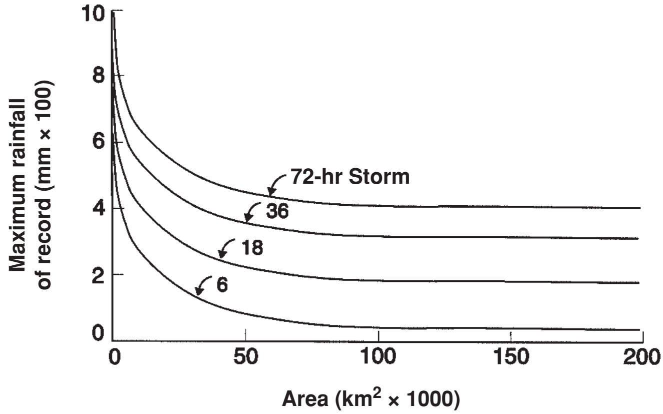


FIGURE 3.10. Relationship between maximum rainfall amounts for specified durations and area (from Hewlett, 1982, © University of Georgia, with permission)

For example, the probability of having a 24-hour rainfall event of 100 mm or greater (Fig. 3.8) over a 20-year period is determined by:

$$\text{Prob}\,(x \geq 100) = 0.025$$

or:

$$\text{Prob}\,(x \geq 100\,\text{mm over 20 years}) = 1 - (1 - 0.025)^{20} = 0.40, \text{ or } 40\%$$

Depth–Area–Duration Analysis

Frequency analysis is based on rainfall characteristics at a point or a specific location. However, the design of storage reservoirs and other water resource-related activities requires that the watershed area be taken into account. Because rainfall does not occur uniformly, it is expected that as larger watersheds are considered, the depth of rainfall associated with a specified probability will decrease (Fig. 3.10). Furthermore, on a given area, the greater the depth of rainfall for a specified duration, the lower the probability of equaling or exceeding that amount.

SNOWFALL

Melting of snowpacks on upland watersheds is often a primary source of water for downstream users in much of the western United States and many other temperate regions of the world. Streamflows originating from snowmelt are not always beneficial to downstream communities, however, as damaging flooding can occur when a large snowpack melts too quickly. Therefore, people living in upland and downstream areas affected by snowmelt are concerned about the magnitude of snowpacks and their rate of melt. Those involved with managing natural resources in uplands and those managing storage reservoirs and water conveyance systems in downstream areas are directly concerned with the measurement of snow and the rate at which an accumulated snowpack will melt. Such issues are related to the science of snow hydrology (see Box 3.2).

Box 3.2

Terminology Used in Snow Hydrology

- Snowpack – mixture of ice crystals, air impurities, and liquid water if melting
- Snowpack density – weight per unit volume; the density of pure ice is 0.92 g/cm^3 and it can vary from $<$0.10 g/cm^3 to $>$0.40 g/cm^3
- Snow water equivalent (WE) – weight of snow expressed as the depth of liquid water over a unit of area = density (ρ) $\times$ depth (D)
- Cold snow – snow with a temperature below 0°C
- Temperature deficit (T_s) – snowpack temperature below 0°C
- Thermal deficiency – heat required to raise the temperature of 1 cm WE of cold snow by 1°C, the heat capacity of water is 1 cal/g/°C; for ice it is 0.5 cal/g/°C; for air it is 0.24 cal/g/°C
- Liquid-water-holding capacity – analogous to soil moisture, it is the water held against gravity on snow crystals and the capillary channels in a snowpack (f); it varies with density, crystal size and shape, and capillarity; $f = 0.03$ on average, generally less than 0.05; at 0°C, $f = 1 - B$ (B is the thermal quality defined below)
- Ripe snowpack – a snowpack that has reached its maximum liquid-water-holding capacity; a snowpack is isothermal at 0°C and it is primed to transmit liquid water
- Cold content (H_c) – heat required to raise the temperature of a cold snow layer of depth D to 0°C; $H_c = 0.5\rho D T_s$
- Latent heat of fusion – heat required to change 1 g of dry snow at 0°C to a liquid state without changing the temperature or from a liquid to solid at 0°C = 80 cal/g = 80 cal/cm^3
- Water equivalent of cold content $W_c = (WE)T_s/160$
- Thermal quality (B) – the ratio of heat required to melt snow to the heat required to melt an equal mass of pure ice at 0°C (B); for a wet, melting snow at 0°C, $B < 1$; for a dry snow at 0°C, $B = 1$; for a cold, dry snow $B > 1$

Methods of Snowfall Measurement

Snowfall can be measured with the standard gauges, but it is more difficult to obtain reliable measurements than for rainfall. Snowfall is more susceptible to wind than rainfall because of its low density and high surface area, which results in rain gauges catching less than the true snowfall. In addition, snow tends to bridge over the orifice of a rain gauge and cling to the sides of containers. If snowfall is to be measured in a standard gauge, the funnel and cylinder are removed and a carefully measured volume of antifreeze is put in the canister. Snow falling into the canister melts in the antifreeze. The difference between the volume of snowfall plus antifreeze and the initially measured volume of the antifreeze represents a measurement of snowfall.

Since snow often remains on the ground for some time, it does not need to be measured as it falls. The most common method of determining the magnitude of a snowpack is to measure its depth on the ground and, if possible, its weight. Such measurements can be made adjacent to standard gauges or they can be obtained in snow surveys over a watershed in which several measurements are taken to estimate the mean snowfall on an area basis.

Snow Surveys

Snow surveys are used to estimate the amount of water in the snowpack and the condition of the snowpack such as its melting potential during the periods of accumulation and melt on courses with a few to 20 or more sampling points along a transect. Snow depth and snow water equivalent (*WE*) are measured at the sampling points on a course by using cylindrical tubes with a cutting edge. These *snow tubes* are graduated in inches or centimeters on the outside of the tube to measure depth. Large-diameter tubes (7.6 cm) are used in areas with shallow snowpacks. Smaller diameter snow tubes (3.8 cm), such as the Mount Rose sampler, are preferable for measuring snowpacks that exceed 1 m in depth.

The manner in which a snow course is laid out depends mostly on how the survey information obtained is to be used. If the purpose of a snow survey is to provide an index of snowpack water equivalent to predict subsequent snowmelt volumes, then regression methods are used. The course need not be representative of the average snowpack for the watershed in question. Instead, a snow course should be located in an accessible area that is relatively flat, is protected from wind, and retains a deep snowpack even in the drier years. The volume of snowmelt is then estimated with a regression relationship based on selected snow course data. The simple regression equations developed are of the general form:

$$\text{Spring snowmelt runoff volume} = PI - LI \tag{3.11}$$

where *PI* is the index of precipitation inputs such as snow water equivalent, spring rainfall, or fall rainfall and *LI* is the index of losses, for example, evapotranspiration (*ET*) estimates.

Multiple regression equations are often developed that have the following components:

$$Y = a + b_1X_1 + b_2X_2 + b_3X_3 - b_4X_4 + \cdots + b_nX_n \tag{3.12}$$

where Y is the volume of snowmelt runoff; X_1 is the maximum snow water equivalent based on snow survey data for April 1, for example; X_2 is the fall precipitation (October to November rainfall); X_3 is the spring rainfall (April); X_4 is the pan evaporation for October to April; and b_i is the regression coefficient.

Multiple regressions such as Equation 3.12 are used widely by the Natural Resources Conservation Service (NRCS) in the western United States.

If the purpose of a snow survey is to provide an estimate of the mean depth of snowpack water equivalent over a watershed, the snow course should be designed to obtain an estimate of the spatial distribution of a snowpack over the watershed area. The same principles that were discussed earlier in estimating the mean depth of rainfall on a watershed should be applied to estimate the mean depth of the snowpack water equivalent on the watersheds.

Snow surveys are often conducted during the periods that normally have maximum snow accumulations, for example, in the western United States from early March through

Box 3.3

Forecasting Future Water Supplies from a Snow Survey

Information obtained from a snow survey can be used to forecast future water supplies. For example, the snow survey administered by the NRCS provides information on the depth and water equivalent of snowpacks at more than 1200 sites in the western states and Alaska to individuals, organizations, and state and federal agencies. This information then becomes a basis to forecast annual water availability, spring snowmelt runoff, and summer streamflows for making decisions relating to flood control, power generation, agricultural crop production, municipal and industrial water supplies, and other water-related uses. Further information on forecasting water supplies from the information obtained in this snow survey is available on the NRCS website (http://www.nrcs.usda.gov/partnerships/links_wsfs.html).

late April. A snow survey can also be conducted at other times in the period that a snowpack is accumulating on a watershed to accommodate other hydrologic objectives such as forecasting future water supplies (Box 3.3).

Remote Sensing Methods

Availability of telemetry and satellite systems has expanded the methods of measuring snowpacks. Telemetry has been particularly useful in transmitting data collected by pressure pillows, called *snow pillows*, located at remote sites. A snow pillow consists of metal plates that sense the weight of the overlying snowpack usually by means of pressure transducers. The weight of the snowpack is then converted into units of snowpack water equivalent. A network called the Snow Data Telemetry System (SNOTEL), operational in the United States, uses radio telemetry to transmit snow pillow data, temperature, and other climatological data from remote mountain areas to processing centers. The data obtained from SNOTEL sites complement snow course data and, importantly, have the advantage of representing current snowpack conditions on the watershed. Hydrologists or watershed managers can retrieve data from SNOTEL any time they wish. For example, the conditions that might lead to a rapid snowmelt can be detected as they evolve and, in doing so, allow for real-time predictions of snowmelt runoff. Additional information about this telemetry system for transmitting snowpack conditions can be found on the SNOTEL website (www.wcc.nrcs.usda.gov/snow/about.html).

Reliable snowpack water equivalent measurements are provided in real time to the NWS offices upon request from NWS hydrologists through the Airborne Snow Survey Program (Carroll, 1987). The airborne measurement technique uses the attenuation of natural terrestrial gamma radiation by the mass of a snow cover to obtain airborne measurements of snow water equivalent often by fixed-wing aircraft. The gamma radiation flux near the ground originates primarily from natural radioisotopes in the soil. In a typical soil, 96% of the gamma radiation is emitted from the upper 20 cm of soil. After a background measurement of radiation and soil moisture is made over a specific flight line, that is, a measurement in the absence of a snow cover, the attenuation of the radiation signal due to the snowpack overburden is used to integrate the average areal amount of water in the snow cover over the flight line. Flight-line surveys have been established for operational streamflow forecasting by the NWS with a network of more than 1900 flight lines in 29 states and 7 Canadian provinces. More information on the NWS Airborne Snow Survey Program is available on the website for the program (www.nohrsc.noaa.gov/snowsurvey/).

Satellite imagery can be useful to assess the extent of snow-covered areas, particularly for large river basins. Resolution of the sensor is used to determine the minimum basin sizes for various satellite systems. Hydrologists and watershed managers in the United States use NOAA-AVHRR (Advanced Very High Resolution Radiometer) data on basins as small as 200 km^2, Landsat MSS (Multispectral Scanner Systems) data on basins as small as 10 km^2, and Landsat TM (Thermal Mapper) data on basins as small as 2.5 km^2 (Rango, 1994). More important than the resolution is the frequency of coverage. The frequency of coverage for many applications is adequate only with the NOAA-AVHRR data, which represents one visible overpass of an area a day. Cloud cover is always a potential problem. However, estimates of snow cover under a partial cloud cover can often be obtained by extrapolation from a cloud-free portion of the basin if available. Relationships are sometimes developed between the percentage of area in a watershed covered with snow and subsequent streamflow due to snowmelt. As applications of satellite technology continue to expand, there will be opportunities for relating many types of spectral characteristics to snowpack condition, melt, and other related processes.

Analysis of Snowfall Measurements

Snowfall data are analyzed for many of the same purposes as rainfall data (discussed above). Long-term representative snowfall data from watersheds or river basins are valuable for water resource managers, streamflow forecasters, and planners to understand current conditions relative to conditions of the past. To improve the accuracy and usefulness of snow data can require one or more of the following:

(1) Missing snowpack measurements on a course can be estimated by regression or other relationships with snowpack measurements from nearby courses.
(2) The consistency of measurements obtained on one snow course relative to the measurements at another course can be determined with double mass analysis.
(3) The recurrence interval of snowfall amounts and the resulting snowmelt runoff can be estimated and compared through frequency analysis.

Snow Accumulation and Metamorphism

The factors affecting the distribution of rainfall on a watershed also affect the deposition of snow on the watershed. However, snowfall on a site can be quite variable on a microscale because its deposition is influenced by wind, topography, forest overstory vegetation, and physical obstructions such as fences. Newly fallen snow usually has a low density, often assumed to be 0.1 g/cm^3, which makes the snowpack susceptible to wind action. It also has a high albedo, ranging from 80% to 95%. Between snowfall and subsequent snowmelt runoff, several changes take place within the snowpack. These changes, called *snowpack metamorphism*, are largely the result of energy exchange and involve changes in the snow structure, density, temperature, albedo, and liquid-water content.

A snowpack undergoes many changes from the time snow falls until snowmelt occurs. Snow particles, which are initially crystalline in shape, become more granular as wind, solar, and sensible-heat energy, and liquid water alter the snowpack. Snow crystals become displaced as the snowpack settles, resulting in an increase in density. Alternate thawing and freezing at the surface of a snowpack followed by periods of snowfall can cause ice planes or lenses to form within the pack. In addition, the temperature of the snowpack changes in response to long periods of either warm or cold weather. When warmer weather dominates, there is a progressive warming of the snowpack but never in excess of a temperature of 0°C. The albedo of snow diminishes over the time that the snowpack is exposed to atmospheric deposition of forest litter, dust, and rainfall. As stated earlier, a new snowpack can have an albedo in excess of 90%. An older snowpack can have an albedo value of less than 45%.

The above discussion provides a qualitative sense of what snowpack metamorphism entails. However, what is needed is a way to quantify snowpack metamorphism to determine when a snowpack is ready to melt and yield liquid water. A snowpack is considered *ripe* when it is primed to release liquid water, that is, when the temperature of the snowpack is 0°C and the liquid-water-holding capacity of the snowpack has been reached. Any additional input of either energy or liquid water to the snowpack will then result in a corresponding amount of liquid water being released from the bottom of the pack.

Cold Content. The energy needed to raise the temperature of a snowpack to 0°C per unit area is called the *cold content*. It is often convenient to express the cold content as the equivalent depth of water entering the snowpack at the surface as rain that upon freezing will raise the temperature of the pack to 0°C by releasing the latent heat of fusion 80 cal/g. Taking the specific heat of ice to be 0.5 cal/g/°C, the following relationship is obtained (Box 3.4):

$$W_c = \frac{\rho D T_s}{160} = \frac{(WE)T_s}{160} \tag{3.13}$$

where W_c is the cold content as equivalent depth of liquid water (cm); ρ is the density of snow (g/cm^3); D is the depth of snow (cm); WE is the snow water equivalent (cm); and T_s is the average temperature deficit of the snowpack below 0°C.

Liquid-Water-Holding Capacity. Like soil, a snowpack can retain a certain amount of liquid water. This liquid water occurs as hygroscopic water, capillary water, and gravitational water. The liquid-water-holding capacity of a snowpack (W_g) is calculated by

$$W_g = f\,(WE + W_c) \tag{3.14}$$

where f is the hygroscopic and capillary water held per unit mass of snow after gravity drainage, usually varying from 0.03 to 0.05 g/g; WE is the snow water equivalent (cm); and W_c is the cold content (cm).

Box 3.4

Cold Content Calculation for a Snowpack of Depth = 100 cm, Snow Density = 0.1 g/cm^3, Temperature Deficit (T_s) = −10°C

Since the heat of fusion equals 80 cal/g and the specific heat of ice equals 0.5 cal/g/°C, the amount of energy required to bring the temperature of a snowpack up to 0°C can be determined by the following sequence of calculations:

$$WE = (0.1\ \text{g/cm}^3)(100\ \text{cm}) = 10\ \text{g/cm}^2$$

To raise the temperature of the pack (which at −10°C would be solid ice) by just 1°C would require:

$$(10\ \text{g/cm}^2)(0.5\ \text{cal/g/°C}) = 5\ \text{cal/cm}^2/\text{°C}$$

To raise the temperature of the snowpack from −10°C to 0°C then would require:

$$[0\text{°C} - (-10\text{°C})](5\ \text{cal/cm}^2/\text{°C}) = 50\ \text{cal/cm}^2$$

To express the energy requirement in terms of equivalent inches of snowmelt:

$$(50\ \text{cal/cm}^2)(1\ \text{g/80 cal})(1\ \text{cm}^3/\text{g}) = 0.63\ \text{cm}$$

Cold content could have been calculated directly by Equation 3.13:

$$W_c = \frac{(0.1)(100)(10)}{160} = 0.63\ \text{cm}$$

Total Retention Storage. The total amount of melt or rain that must be added to a snowpack before liquid water is released is the total retention storage (S_f):

$$S_f = W_c + f\,(WE + W_c) \qquad \textbf{(3.15)}$$

The snowpack is said to be *ripe* when the total retention storage is satisfied.

Snowmelt

Once the snowpack is ripe, any additional input of energy will result in meltwater being released from the bottom of the snowpack. The main sources of energy for snowmelt to occur are essentially the same as those required for *ET*. To calculate snowmelt, therefore, requires application of either an energy budget approach or the empirical approximation of the energy available for snowmelt.

Energy Budget and Snowmelt Relationships

The energy that is available either to ripen a snowpack or to melt snow can be determined with an energy budget analysis. The energy budget can be used to determine the available energy for ripening or melting as follows:

$$M = W_{\mathrm{i}}(1 - \alpha) + W_{\mathrm{a}} - W_{\mathrm{g}} + H + G + LE + H_{\mathrm{r}} \tag{3.16}$$

where M is the energy available for snowmelt (cal/cm^2); W_i is the total incoming shortwave (solar) radiation (cal/cm^2); α is the albedo of snowpack (fraction); W_a is the incoming longwave radiation (cal/cm^2); W_g is the outgoing longwave radiation (cal/cm^2); H is the convective transfer of sensible heat at the snowpack surface (cal/cm^2), which can be + or – as a function of gradient; G is the conduction at the snow–ground interface (cal/cm^2); LE is the flow of latent heat (condensation [+], evaporation or sublimation [–]) (cal/cm^2); and H_r is the advective heat from rain or fog (cal/cm^2).

Snowmelt can be determined directly if all the above energy components can be measured. The amount of snow that melts from a specific quantity of heat energy depends on the condition of the snowpack which can be expressed in terms of its thermal quality. *Thermal quality* is the ratio of heat energy required to melt 1 g of snow to that required to melt 1 g of pure ice at 0°C, and is expressed as a percentage (Box 3.5). A thermal quality less than 100% indicates the snowpack is at 0°C and contains liquid water. Conversely, a thermal quality greater than 100% indicates the snowpack temperature is less than 0°C with no liquid water. Snowmelt, therefore, can be determined by:

$$M = \frac{\text{Total Energy (cal/cm}^3)}{B \text{ (80 cal/g)}} \tag{3.17}$$

where M is the snowmelt (cm) and B is the thermal quality expressed as a fraction.

Under most conditions, however, total energy cannot be measured. Therefore, approximations of this energy (M), such as obtained by applications of the generalized snowmelt equations or the temperature index (degree-day) method, are used to estimate snowmelt.

Generalized Snowmelt Equations

The generalized snowmelt equations were developed from extensive field measurements at snow research laboratories throughout the western United States (USACE, 1956, 1998). All of the major energy components considered in their development are outlined in the following discussion by considering each major source of energy separately.

Solar Radiation. Solar radiation is the major source of energy for snowmelt. The amount of solar radiation reaching a snowpack surface is dependent upon the slope and aspect of the surface, cloud cover, and forest cover. In the Northern Hemisphere, south-facing slopes receive more radiation than north-facing slopes. The more moderate the slope, the more moderate the effect that the slope has on solar radiation. The corresponding potential solar radiation for a specified latitude can be determined by knowing the slope and aspect of an area.

The amount of solar radiation striking the surface of a snowpack in the open is a function of the percentage of cloud cover and the height of the clouds (Fig. 3.11). The percentage of incoming solar radiation that reaches the snow surface on a forested watershed depends largely on the type of cover, density, and condition of the forest canopy. As the density

Box 3.5

Determining Snow Thermal Quality by the Calorimeter Method (Snow Depth = 60 cm, Volume of Snow Sample Taken = 15,000 cm^3, and Weight of Sample = 2000 g) (modified from Hewlett 1982)

The above sample was placed in a thermos which initially contained 7000 g of water at a temperature of 32°C. After adding the snow sample, and after all the snow melted, the temperature of the water in the thermos was 8°C. The following calculations were made:

$$\text{Snow density} = \frac{2000 \text{ g}}{15{,}000 \text{ cm}^3} = 0.133 \text{ g/cm}^3$$

$$\text{Snow water equivalent (WE)} = (0.133)(60 \text{ cm}) = 8.0 \text{ cm}$$

First, determine the heat energy needed to raise the temperature of the melted snow water from 0°C to 8°C:

$$(2000 \text{ g})(8°\text{C})(1 \text{ cal/g/°C}) = 16{,}000 \text{ cal}$$

The heat available energy was:

$$(7000 \text{ g})(32 - 8°\text{C})(1 \text{ cal/g/°C}) = 168{,}000 \text{ cal}$$

The heat available to melt ice would equal:

$$168{,}000 \text{ cal} - 16{,}000 \text{ cal} = 152{,}000 \text{ cal}$$

Therefore,

$$\text{Ice content} = \frac{152{,}000 \text{ cal}}{80 \text{ cal/g}} = 1900 \text{ g}$$

$$\text{Thermal quality } (B) = \frac{1900 \text{ g}}{2000 \text{ g}} = 0.95 = 95\%$$

The snowpack contained 5% liquid water and was at a temperature of 0°C.

of a forest canopy increases, incoming solar radiation decreases exponentially (Fig. 3.12). As a result, with dense canopies, the effect of forest cover overrides the effects of cloud cover. Coniferous forest cover can reduce substantially the amount of solar radiation that reaches a snow surface, while deciduous forests have less of an effect on solar radiation. For either open or forested conditions, the solar radiation (W_i) reaching the snowpack surface is further reduced by the snowpack albedo (α). Snowmelt from solar radiation (M) is then defined as

$$M = \frac{W_i(1 - \alpha)}{B(80)} \tag{3.18}$$

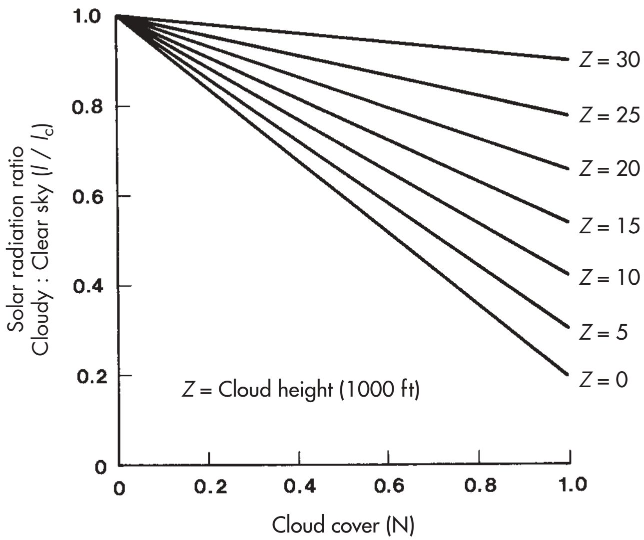


FIGURE 3.11. Relationship between solar radiation and cloud height and cloud cover (from USACE, 1956)

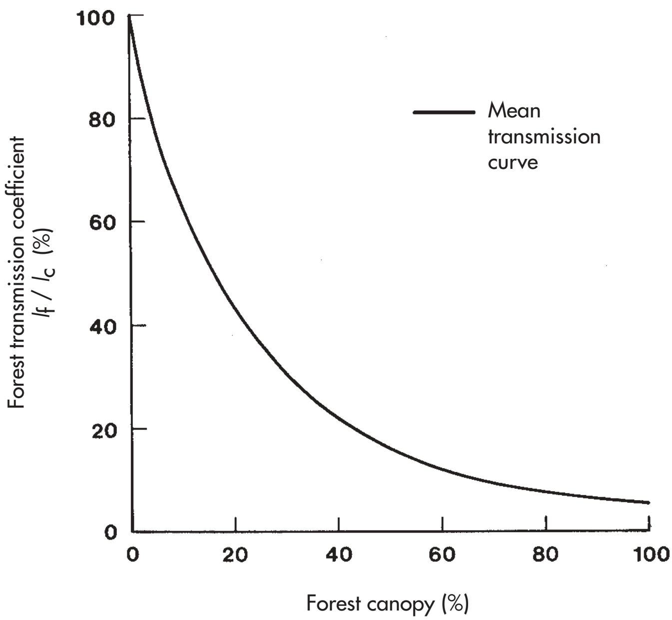


FIGURE 3.12. Relationship between coniferous forest canopy density and transmission of solar radiation (from USACE, 1956)

Longwave Radiation. Snow absorbs and emits nearly all of the incident longwave or terrestrial radiation. Net longwave radiation at a snow surface is determined largely by overhead back radiation from the earth's atmosphere, clouds, and forest canopies.

Under clear skies, back radiation from the atmosphere is a function of the water content of air. Over snowpacks, the vapor pressure of air usually varies between 3 and 9 mb pressure, resulting in a fairly constant rate of longwave radiation of 0.757 cal/cm^2/min. The narrow range of snowpack temperatures limits the amount of radiation emitted from a snowpack. Since a snowpack temperature cannot exceed 0°C, the maximum longwave radiation a ripe snowpack can emit is 0.459 cal/cm^2/min. Therefore, under clear skies, the net longwave radiation at a snow surface is approximately:

$$R_c = 0.757\sigma T_a^4 - 0.459 \tag{3.19}$$

where R_c is the net longwave radiation (cal/cm^2/min); σ is the Stefan–Boltzmann constant, which is equal to 0.826×10^{-10} (cal/cm^2/min/K^4); and T_a is the air temperature (K).

When clouds are present, the temperature at the base of the clouds determines the back radiation to the snow surface, and the net longwave radiation becomes:

$$R_{cl} = \sigma\left(T_c^4 - T_s^4\right) \tag{3.20}$$

where R_{cl} is the net longwave radiation under cloudy skies; T_c is the temperature at the base of clouds (K); and T_s is the snowpack temperature (K).

Similarly, the net longwave radiation (R_f) for a snowpack beneath a dense conifer canopy is:

$$R_f = \sigma\left(T_f^4 - T_s^4\right) \tag{3.21}$$

where T_f is the temperature of the underside of the forest canopy (K).

Because temperatures of the trees are rarely known, T_f is often estimated from air temperature (T_a). Under clear-sky conditions and a ripe snowpack, the net longwave radiation under a conifer forest canopy is:

$$R_{lw} = \sigma T_a^4\,[F + 0.757(1 - F)] - 0.459 \tag{3.22}$$

where F is the canopy density expressed as a decimal fraction.

Net Radiation. The net (all-wave) radiation that is available for snowmelt is governed largely by forest-cover conditions and cloud conditions. For conditions of little or no forest cover, clouds strongly affect the net radiation at the snowpack surface. Under partial or dense forest conditions, forest cover governs net radiation at the snowpack surface.

There is a tradeoff between shortwave and longwave radiation at a snowpack surface as forest cover changes. As forest cover increases, the solar radiation at the snowpack surface is reduced greatly; the longwave radiation loss from the snowpack is reduced; and the longwave gain component from the canopy increases (Fig. 3.13). Net radiation at the snowpack surface is at a minimum between 15% and 30% canopy cover, but increases at dense forest canopy conditions because of the much higher net longwave radiation. Net radiation is highest at 0% cover with large inputs of solar radiation. These relationships have implications for the management of forested watersheds in snow-dominated regions.

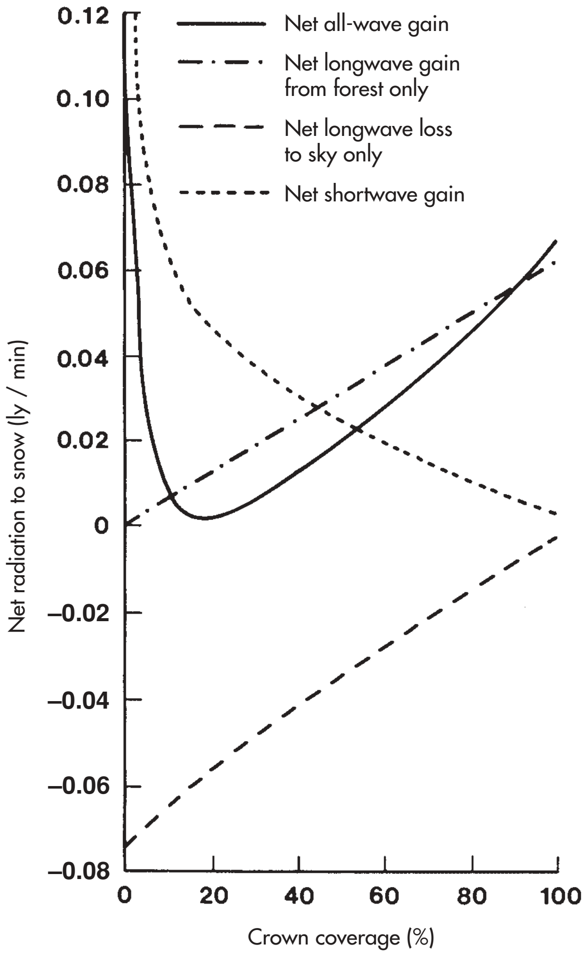


FIGURE 3.13. Net radiation of a snowpack in relation to a forest canopy cover (from Reifsnyder and Lull, 1965)

Convection–Condensation Melt. Significant amounts of heat energy can be added to snowpacks by the turbulent exchange of sensible heat from the overlying air and direct condensation on the snow surface. Some of the largest snowmelt rates, sometimes producing snowmelt floods, occur when warm moist air masses move into a watershed with significant snow cover. In such instances, both sensible heat from turbulent air and condensation on the snowpack surface contribute to high snowmelt rates. In general terms, this energy exchange can be approximated by

$$q_{\mathrm{e}} = A_{\mathrm{e}}\left[\frac{\mathrm{d}q}{\mathrm{d}z}\right] \tag{3.23}$$

where q_{e} is the energy exchange (cal/cm^2/t); A_{e} is the exchange coefficient; and $\mathrm{d}q/\mathrm{d}z$ is the vertical gradient of temperature or water vapor, t is the time interval.

For a ripe snowpack, the snow surface would have a temperature of 0°C with a corresponding vapor pressure of 6.11 mb. Using these values for the snow surface and measurements of air temperature and relative humidity of the air above the snow surface, the melt caused by the energy exchange from convection and condensation can be estimated.

Rain Melt. Snowmelt caused by the addition of sensible heat from rainfall to a snowpack is relatively small in magnitude. It can be estimated from air temperature and rainfall measurements for a snowpack with a thermal quality of 100% as

$$M_{\mathrm{p}} = \frac{P_{\mathrm{r}} T_{\mathrm{a}}}{80\ \mathrm{cal/g}} = 0.013 P_{\mathrm{r}} T_{\mathrm{a}} \tag{3.24}$$

where M_{p} is the daily snowmelt (cm); P_r is the daily rainfall (cm); and T_{a} is the average daily air temperature (°C).

Again, the above equation holds for a ripe snowpack. If the snowpack has a temperature below 0°C, additional energy can be released to the pack by virtue of the release of the heat of fusion, which is 80 cal for every 1 cm of rain that freezes.

Conduction Melt. Heat energy can be added to the base of a snowpack by conduction from the underlying ground. For any day, the amount of energy available by conduction and, hence, the amount of melt are relatively small. Daily values of 0.5 mm are frequently assumed, and sometimes this component is simply ignored if snowmelt is being estimated over a short period. However, conduction melt (ground melt) can be significant for seasonal estimates of snowmelt. For example, monthly snowmelt totals in the central Sierra Nevada in California ranged from 0.3 cm in January to 2.4 cm in May (USACE, 1956).

Combined Snowmelt Equations. The energy–snowmelt relationships derived above provide the foundation for a set of generalized equations that can be used to estimate snowmelt for forested watersheds (USACE, 1998). Because of the influence of a forest cover on energy exchange, the forest-cover condition of a watershed determines which equation to use. For rain-free periods, the daily melt from a ripe snowpack (isothermal at 0°C, with a 3% liquid-water content) is calculated by one of the following equations, which are differentiated on the basis of the forest canopy cover (F):

Heavily Forested Area ($F > 0.80$):

$$M = 0.19\, T_{\mathrm{a}} + 0.17\, T_{\mathrm{d}} \tag{3.25}$$

Forested Area ($F = 0.60$–0.80):

$$M = k(0.00078\, v)(0.42\, T_{\mathrm{a}} + 1.51\, T_{\mathrm{d}}) + 0.14\, T_{\mathrm{a}} \tag{3.26}$$

Partly Forested Area ($F = 0.10$–0.60):

$$M = k'(1 - F)0.01\, W_{\mathrm{i}}(1 - \alpha) + k(0.00078\, v) \times (0.42\, T_{\mathrm{a}} + 1.51\, T_{\mathrm{d}}) + F(0.14\, T_{\mathrm{a}}) \tag{3.27}$$

Open Area ($F < 0.10$):

$$M = k'(0.0125\, W_{\mathrm{i}})(1 - \alpha) + (1 - N)(0.014\, T_{\mathrm{a}} - 2.13) + N(0.013\, T_{\mathrm{c}}) + k(0.00078\, v)(0.42\, T_{\mathrm{a}} + 1.51\, T_{\mathrm{d}}) \tag{3.28}$$

where M is the daily melt (cm); T_a is the air temperature (°C) at 2 m above the snow surface; T_d is the dew point temperature (°C) at 2 m above the snow surface; v is the wind speed (km/day); W_i is the observed or estimated solar radiation (cal/cm^2/day); α is the snow surface albedo (decimal fraction); k' is the basin shortwave radiation melt factor, which depends on the average exposure of the open areas to solar radiation compared to an unshielded horizontal surface; F is the average forest canopy cover a for watershed (expressed as a decimal fraction); T_c is the cloud-base temperature (°C); N is the cloud cover expressed as a decimal fraction; and k is the basin convective–condensation melt factor, which depends on the relative exposure of the watershed to wind.

The only empirically fitted parameters in the above equations are k and k'. The basin shortwave radiation melt factor (k') can be estimated from solar radiation data for a specified latitude, slope, and aspect. Because one has to average several slopes and aspects for a watershed, the k' value represents an average for the watershed. Watersheds with a southern exposure would tend to have a $k' > 1$, while those with northern exposures would have a $k' < 1$. A watershed that has a balance between north- and south-facing slopes would have a $k' = 1.0$. The basin convective–condensation melt factor (k) is similarly an average value for a particular watershed that indicates the exposure of the snowpack to wind. These values range from $k = 1.0$ for open areas to $k = 0.8$ for dense forest cover.

Temperature Index (Degree-Day) Method

The data that are required to apply the generalized snowmelt equations can restrict the use of these equations in many situations, such as in remote areas and for day-to-day operational conditions. As a result, a method that depends solely on knowledge of air temperature can be used to estimate snowmelt.

The temperature index method is based on the following empirically derived equation of the form:

$$M = MR\,(T_a - T_b) \tag{3.29}$$

where M is the daily snowmelt (cm); MR is the melt-rate index or degree-day factor (cm/°C/day); T_a is the daily air temperature value, usually either the average daily or the maximum daily temperature (°C); and T_b is the base temperature, at which no snowmelt is observed (°C).

The difference T_a–T_b represents the degree-days of heat energy available for melting a snowpack. The melt-rate index relates this heat energy to snowmelt on the watershed. Although we know that the air temperature is but one source of energy for snowmelt, it correlates well with radiation inputs, particularly in the snowmelt season, and it is also a reasonable index for forested watersheds.

The equation for the temperature index method (Equation 3.29) is derived from regression analysis of daily snowmelt versus air temperature (Fig. 3.14). The melt-rate index is the slope of the regression line, and the base temperature (T_b) is that air temperature at which no melt is observed. Examples of temperature index coefficients for three watersheds are presented in Table 3.2.

Since the temperature index method is developed for a specified watershed, both the melt-rate index and the base-temperature values are likely to differ from one watershed

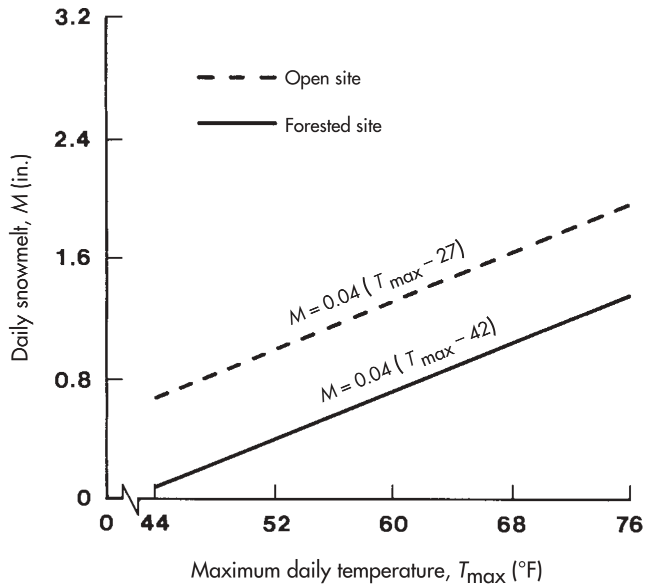


FIGURE 3.14. Temperature index relationships for maximum daily air temperatures (from USACE, 1956)

to the next. Furthermore, the degree-day factor or melt-rate index can vary seasonally rather than remain constant as often assumed (Rango and Martinec, 1995). The elevation difference and proximity of the air temperature station to the watershed will affect the values derived. Nevertheless, the temperature index method has been used successfully to predict snowmelt in mountainous watersheds. It has also been shown to be a reliable alternative to the more physically based and data-demanding energy budget method in predicting snowmelt for computer simulation models (Rango and Martinec, 1995).

TABLE 3.2. Base temperature and melt-rate indices for three watersheds in the western United States

	Mean temperature			Maximum temperature		
		Melt-rate index			Melt-rate index	
Watershed	Base T (°F)	April	May	Base T (°F)	April	May
Central Sierra, California	26	0.036	0.062	29	0.020	0.038
Upper Columbia, Montana	32	0.037	0.072	42	0.109	0.064
Willamette, Oregon	32	0.039	0.042	42	0.046	0.046

Source: From USACE (1956).

Summary and Learning Points

Precipitation input to a watershed can be in the form of rainfall or as the amount of melt resulting from snowfall. You should have gained a general understanding of the factors that influence the occurrence of a precipitation, including rainfall, snowfall, and snowmelt. Specifically, you should be able to

(1) Describe the conditions necessary for precipitation to occur.
(2) Explain the different precipitation and storm characteristics associated with frontal storm systems, orographic influences, and convective storms.
(3) Understand how rainfall can be measured at a point and how these measurements can be used to estimate the average depth of rainwater over a watershed area.
(4) Estimate the values of rainfall that are missing for a storm.
(5) Explain the purpose of performing double mass analysis and frequency analysis.
(6) Understand how snowfall can be measured and the purpose of a snow survey.
(7) Explain the causal factors affecting snow accumulation, snow ripening, and snowmelt.
(8) Explain and be able to quantify the snow-ripening process of a snowpack and explain the effects of snowpack condition on spring snowmelt floods.
(9) Describe the differences between snowmelt under a conifer forest and that in an open field using an energy budget approach.
(10) Calculate snowmelt using either the appropriate generalized snowmelt equation or the temperature index method and discuss the advantages and disadvantages of these two methods.

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CHAPTER 4 Evaporation, Interception, and Transpiration

INTRODUCTION

Evaporation from soils, water bodies, and plant interception and transpiration are considered collectively as *evapotranspiration* (*ET*). *ET* affects water yield, largely determines what proportion of precipitation input to a watershed becomes streamflow, and is influenced by land-use activities that alter vegetation and water bodies on the landscape. The water budget equation can be used to estimate *ET* over a period of time as follows:

$$ET = P - Q - \Delta S - \Delta l \qquad \textbf{(4.1)}$$

where *ET* is the evapotranspiration (mm); *P* is the precipitation (mm); *Q* is the streamflow (mm); ΔS is the change in the amount of storage in the watershed, $S_2 - S_1$ (mm), where S_2 is the storage at the end of a period and S_1 is the storage at the beginning of a period; and Δl is the change in deep seepage, $l_o - l_i$ (mm), where l_o is the seepage out of the watershed and l_i is the seepage into the watershed.

The *ET* component of the water budget is more than 95% of the 300 mm of annual precipitation in Arizona. It is more than 70% of the annual precipitation for the entire United States (Gay, 1993). The ratio *ET*/*P* is close to unity for dry climates, meaning that the ratio *Q*/*P* is small. *ET*/*P* is less in humid climates where the magnitude of *ET* is governed by available energy rather than the availability of water. Changes in vegetation that reduce *ET* will increase streamflow and/or groundwater recharge while increases in *ET* have the opposite effect.

Rates of *ET* influence water yield by affecting the antecedent water status of a watershed. High rates deplete water in the soil and in surface water impoundments, leaving more space that is available to store precipitation. Low *ET* rates leave less storage space in the soil

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

and in surface water impoundments. The amount of storage space in a watershed affects the amount and, to some extent, the timing of streamflow resulting from precipitation events.

ET is the result of cumulative evaporation processes with each process requiring a change in the state of water from liquid to vapor and the net transfer of this vapor to the atmosphere. Therefore, there must be a flow of energy to the evaporating or transpiring surface before evaporation or transpiration can occur. Evaporation occurs at the water surface in the case of lakes or ponds. However, interception is a more complicated process because evaporation occurs from wet plant surfaces, plant residues, and vegetative litter.

Transpiration (*T*) is the most complex evaporative process. It requires a flow of liquid water to plant cell surfaces in leaves and exits through plant stomata in most plants. The rate of evaporation depends on the rate of vapor flow away from these surfaces in all instances. If one or more of these flows are changed, there is a corresponding change in the total evaporative loss from a surface. The flow of energy to evaporating surfaces has been described in Chapter 2 and will be discussed in terms of potential evapotranspiration (*PET*) in this chapter.

The Evaporation Process

ET requires both energy and conditions that permit water vapor to flow away from evaporating or transpiring surfaces. Water molecules migrate from the liquid surface as a result of their kinetic energy. This transfer involves a change of state from liquid to vapor caused by energy inputs to the evaporating surface.

Vapor flow is initially a *diffusion process* in which water molecules diffuse from a region of higher concentration (the evaporating surface or source) toward a region of lower concentration (a sink) in the atmosphere. Water molecules at the soil–atmosphere or leaf–atmosphere interface must first diffuse through the *boundary layer*. This is also the layer through which sensible heat is transferred by molecular conduction only. The boundary layer of air adjacent to evaporating and transpiring surfaces can be as thin as 1 mm or less, but is at maximum thickness under still-air conditions. Wind and air turbulence reduce the boundary layer thickness but there is no turbulent flow in the boundary layer itself.

Once water molecules leave the boundary layer, they move into a turbulent zone of the atmosphere where further movement is primarily by *mass transport*, that is, turbulent eddy movement. In mass transport, whole parcels of air or eddies with water vapor and sensible heat flow in response to atmospheric pressure gradients, which cause the air parcels to flow both vertically and horizontally.

Evaporation describes the *net* flow of water away from a surface. Water molecules also return to the evaporating surface by mass transport and diffusion processes. If the amount of vapor arriving equals the amount leaving, a steady state exists and no evaporation occurs. If more molecules arrive than leave a surface, a net gain results, which is *condensation*.

The vapor pressure of water molecules at the evaporating surface must exceed the vapor pressure in the atmosphere for evaporation to occur. Under natural circumstances, the vapor pressure of liquid water is mainly a function of its temperature although solute content, atmospheric pressure, and water-surface curvature in capillaries can also be important. The vapor pressure of water molecules in the atmosphere is primarily a function of air temperature and humidity of the air (see Fig. 3.1). The vapor pressure gradient between evaporating surfaces (e_s) and the atmosphere (e_a) is the driving force that causes a net movement of water molecules. This vapor pressure deficit ($e_s - e_a$) between an evaporating surface and the atmosphere (see the difference between points A and B in Fig. 3.1) is often a component of empirical equations that are described later.

Conceptual relationships of the evaporative process have been developed for complex surfaces such as plants and soils. Such models assume that the vapor flow away from evaporating (E) or transpiring (T) surfaces is directly proportional to the vapor pressure deficit ($e_s - e_a$) and inversely proportional to the resistance (R_v) of air to the molecular diffusion and mass transport of water vapor. That is

$$E \text{ or } T = \frac{e_s - e_a}{R_v} \tag{4.2}$$

The R_v term includes the resistance for the turbulent layer of the atmosphere (r_e) and the boundary layer (r_{bl}) and the internal resistance characterizing air-filled soil pores (r_s) or plant pores (r_{st}) called *stomates*. For convenience, the two atmospheric resistances are sometimes combined as r_a and the two internal resistances as r_n. The internal resistance is necessary because the liquid–air interface is often beneath the external surface of soil or plants. Water vapor diffusing outward encounters resistance from the air in soil and plant pores before reaching the boundary layer. For a wet external surface, the diffusive path length consists only of the boundary and turbulent layers of the atmosphere. The extra path length for dry soil or plants increases the total resistance to flow.

Evaporation from Water Bodies

Evaporation from lakes, ponds, reservoirs, or swimming pools is determined only by energy and vapor flows. Because many areas of the world depend on reservoirs to provide municipal water supplies and water for irrigation, evaporation losses are important in determining whether the reservoir storage is sufficient to meet water demands. Although municipal water supply reservoirs are found in dry and wet regions alike, the greatest evaporation rates tend to occur in the driest regions where water is scarcer (Table 4.1).

Methods of determining lake evaporation include the water budget approach or the use of empirical relationships. Lake evaporation can be estimated with the water budget

Table 4.1. Evaporation from selected lakes and reservoirs in the United States and Australia

Location	Annual evaporation (mm)
Lake Superior (1959–1966)	760
Pyramid Lake, Nevada	1118
Devils Lake, North Dakota	366
Oklahoma City (reservoir)	1676
Seattle, Washington (reservoir)	610
Salt Lake City, Utah (reservoir)	1397
Lake Eyre, Australia	2134

Source: Data from van der Leeden et al. (1990).

method if all components of the water budget except evaporation are either measured or estimated over a period of time (t):

$$E = Q_i + GW_i + P - Q_o - GW_o - \Delta S \tag{4.3}$$

where E is the lake evaporation (m^3/t); P is the precipitation on the lake surface (m^3/t); Q_i and Q_o are surface water flows into and out of the lake (m^3/t), respectively; GW_i and GW_o are subsurface or groundwater flows into and out of the lake (m^3/t), respectively; and ΔS is the change in storage (m^3/t).

Measuring all of the inflow and outflow components of a lake is difficult. Although it is straightforward to measure surface streams that enter and leave a lake, surface runoff directly into the lake from the surrounding area cannot be directly measured. Neither can the subsurface or groundwater flows be directly measured. In some circumstances, these components can be estimated with the use of either hydrologic simulation models or methods of proportioning surface and groundwater components using stable isotopes (see Chapter 16).

One method of estimating lake evaporation is (Dunne and Leopold, 1978)

$$E_o = N\,(e_s - e_a)\,f\,(u) \tag{4.4}$$

where E_o is the evaporation from the water body (mm); N is the mass transfer coefficient determined empirically; e_s is the vapor pressure of the water surface (mb); e_a is the vapor pressure of the air (mb); and $f(u)$ is the function of wind speed (km/day).

The vapor pressure deficit or gradient ($e_s - e_a$) above a free-water surface can be determined from measurements of surface water temperature, air temperature, and the relative humidity of the air.

Evaporation from lakes or reservoirs is commonly estimated with the *pan evaporation* method. With this approach, evaporation is measured from a pan and a coefficient is applied as follows:

$$PET = C_e E_p \tag{4.5}$$

where C_e is the pan coefficient and E_p is the pan evaporation (mm/day).

The standard evaporation pan in the United States, a National Weather Service Class A pan, is a metal cylinder 122 cm in diameter and 25 cm deep. Water depth is maintained at 18–20 cm and measured daily with a hook gauge in a stilling well.

The pan evaporation method has been used extensively to estimate evaporation from lakes and reservoirs for which C_e usually ranges from 0.5 to 0.8. Average annual pan coefficients of 0.70–0.75 are often used for lakes when they have not been derived experimentally. Seasonal relationships of C_e for lakes tend to be <1.0 for the warm season because the metal pan heats the water at a greater rate than would occur in a natural lake. For deep lakes during cold periods, however, C_e can be >1.0 because the energy that is stored by a natural deep water body can provide energy for evaporation even during periods of cold air temperature. Pan evaporation data have also been used as estimates of *PET* that is described later in this chapter. Values of C_e have been derived to simulate seasonal changes in transpiration for certain plant species in some instances.

Evaporation from Soil Surfaces

Evaporation from a soil is a more complex phenomenon than evaporation from a water body. Given a bare, flat, wet soil surface, the water supply is initially unlimited and the amount of evaporation depends on the *energy supply* and the *vapor pressure gradient* as with a water body. When soils are exposed to the open atmosphere, sufficient vapor pressure gradients usually exist and are maintained, causing only the energy supply to limit evaporation from the wet and exposed soil. If a piece of transparent plastic sheeting were laid over this soil, evaporation would cease because vapor flow would be blocked even though energy and water flows would still converge at the active surface. In natural environments, however, energy inputs to the active surface increase the vapor pressure of water, steepening the vapor pressure gradient.

Evaporation proceeds at rates similar to those of free-water surfaces, assuming equal energy input in this case. As evaporation occurs, the lost water is replaced by water moving up from below the evaporating surface through the connecting water films around soil particles and through capillary pores (refer to Fig. 2.9). As the soil surface dries, the gradient in total water potential ($d\psi/dx$) in the soil increases with water moving through the soil from the zone of higher potential (lower layer, wet zone) to the region of lower potential (upper layer, drier evaporating surface). Water films around the soil particles become thin, and the pathway through which water must move to reach the evaporating surface becomes more tortuous, thereby reducing hydraulic conductivity. After a period of drying, the rate of water flow through the soil limits the rate of evaporation at the soil surface.

Water flow in moist soil is primarily liquid but as soils dry, vapor diffusion through pores becomes more dominant. At about -1500 kPa water potential, water flow occurs mainly as vapor or some combination of vapor and liquid. Liquid flow is reduced greatly at -1500 kPa because the continuity of capillary water and water films becomes disrupted; this is the point at which most plants become wilted referred to as the *permanent wilting point* (see Chapter 2). Water cannot move as rapidly through soils in vapor form as in liquid form. As a consequence, deficient soil moisture limits evaporation at the active surface regardless of the energy input. With time, evaporation rates decrease because the water flow to the active evaporating surface is too slow to keep pace with the energy input. How much water will evaporate from a soil under these conditions depends largely on soil texture. Fine-textured soils retain pore water continuity at lower water contents than coarse-textured soils. Therefore, soil evaporation diminishes sooner in sandy soils than in clay soils, which have smaller pores that permit water films to remain intact for a longer period.

Interception

Once rainfall or snowfall occurs, the type, extent, and condition of vegetation and plant litter influence the pattern of deposition and the amount of precipitation reaching the soil surface. Dense coniferous forests in northern latitudes and the multistoried canopies of the tropics intercept large quantities of precipitation that return directly to the atmosphere by evaporation, becoming a loss of water from a watershed. Interception losses are less in arid and semi-arid environments that have sparser vegetation.

Not all the precipitation caught by a living vegetative canopy is lost to the atmosphere. Much can drip off the foliage or run down the stems of trees, ultimately reaching the

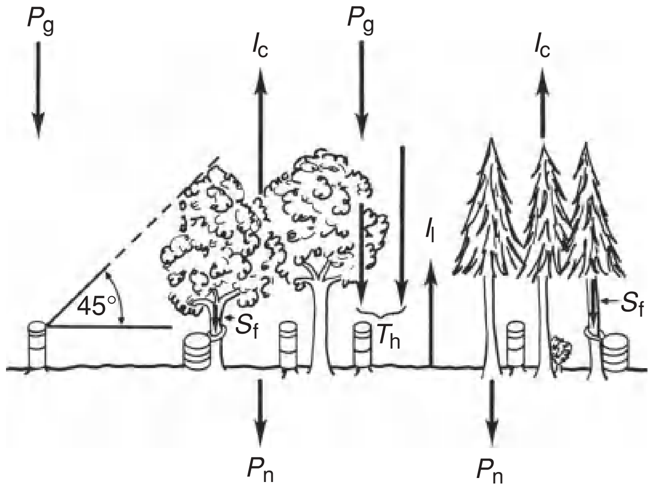


FIGURE 4.1. Components of interception (from Hewlett, 1982 © University of Georgia Press, with permission). I_c, canopy interception loss; I_l, litter interception; P_g, gross precipitation; P_n, net precipitation; S_f, stem flow; T_h, throughfall

soil surface. Plant residue or litter at the soil surface can store part of this water, but once saturated, drainage from plant residue or forest floor litter will reach the mineral soil surface.

Components of Interception

The components of the interception process, methods of measurement, and the resulting deposition of precipitation for a forest canopy are illustrated in Figure 4.1. Interception by the forest canopy is defined as:

$$I_c = P_g - T_h - S_f \tag{4.6}$$

where I_c is the canopy interception loss (mm); P_g is the gross precipitation (mm); T_h is the throughfall, precipitation that passes through the vegetative canopy or as drip from vegetation (mm); and S_f is the stemflow, water that flows down the stems to the ground surface, measured with collars fixed to the stems of trees that divert stemflow to containers for measurement (mm).

The partitioning of a given quantity of rainfall into the above pathways depends on vegetative cover characteristics such as leaf type, leaf and branch surface area, branch attitude, shape of the canopy, and roughness of the bark. The interception components of a growing forest from seedling stage to maturity change as follows:

- T_h diminishes over time as the canopy cover increases.
- S_f increases over time but is always a small quantity.
- Storage capacity of vegetation (leaf surface area) and litter increases.

Throughfall

Canopy coverage, total leaf area, the number of layers of vegetation, and rainfall intensity determine how much of the gross rainfall reaches the forest floor. The size and shape of canopy openings affect the amount, intensity, and spatial distribution of throughfall. The shape of the forest overstory, particularly the branch and leaf angles, can concentrate throughfall in drip points with larger drop sizes, velocities, and kinetic energy than from rainfall in adjacent open areas. Throughfall can be variable with some locations receiving little throughfall.

Relationships between throughfall and precipitation have been developed for forests in many areas of the world (Box 4.1). Because throughfall depends on tree canopy surface area and cover, descriptors of forest stands that relate to these characteristics should be included in prediction equations to allow for wider application.

Box 4.1

Throughfall Relationships for Different Vegetative Canopies

1. Eastern Hardwood Forests, United States (Helvey and Patric, 1965):
   Growing Season

   $$T_h = 0.901\ (P) - 0.031\ (n)$$

   Dormant Season

   $$T_h = 0.914\ (P) - 0.015\ (n)$$

   where T_h is throughfall (in); P is total precipitation (in); and n is number of storms
2. Southern Pine Forests, United States (Roth and Chang, 1981):
   Longleaf Pine

   $$T_h = 1.002\ (P) - 0.0008\ (P)^2 - 1.397$$

   Loblolly Pine

   $$T_h = 0.930\ (P) - 0.0011\ (P)^2 - 0.610$$

   where T_h is throughfall (mm) and P is total rainfall (mm)
3. New Zealand Vegetation Communities (Blake, 1975):
   Kauri Forest

   $$T_h = 0.60\ (P) - 3.71$$

   Manuka Shrub

   $$T_h = 0.44\ (P) - 0.10$$

   Mountain Beech Forest

   $$T_h = 0.69\ (P) - 1.90$$

   where T_h is throughfall (mm) and P is total rainfall (mm)

Stemflow

Stemflow, usually less than 2% of gross annual precipitation, is affected by branch attitude, shape of tree crowns, and roughness of bark. Tree species with rough bark generally retain more water and exhibit less stemflow than those with smooth bark.

Higher stemflows have been reported, however. In Australia, stemflow amounted to 5% and 9% of precipitation for sclerophyll eucalyptus and Monterey pine (*Pinus radiata*) forests, respectively (Crockford and Richardson, 1990). Stemflow varies considerably among eucalyptus species with those having smooth and nonabsorptive bark exhibiting high stemflow amounts while those with thick, fibrous, and absorptive bark exhibiting little stemflow.

Although stemflow is not a large quantity in terms of an annual water budget, the process can be an important mechanism of replenishing soil moisture. Stemflow concentrates water in a small area near the base of the tree stem; this water can flow quickly and penetrate deeply into the soil through root channels.

Interception Process

The type of precipitation, whether rain or snow, the intensity and duration of rainfall, wind velocity, and evaporative demand affect interception losses. While clearly visible in a conifer forest immediately after snowfall, the interception of snow might not represent a significant loss of water in some cases (Box 4.2). However, interception of snow has been shown to significantly affect annual water yield in the subalpine forests of the Rocky Mountains of the United States. Approximately 50% of water yield increases following removal of

Box 4.2

Deposition of Intercepted Snowfall

Photograph records in Colorado (Hoover and Leaf, 1967) and Arizona (Tennyson et al., 1974) inferred that intercepted snowfall remains in coniferous trees for a short time period. Much of the intercepted snow eventually reaches the ground because of wind erosion and snowmelt with subsequent dripping and freezing in the snowpack. While snowfall interception by trees might not be a significant loss to the small-scale water budget surrounding individual trees, the deposition of intercepted snow on the ground results in a redistribution of the snowpack. On a much larger scale, it appears that both small patch cuts and thinning in subalpine forests on Fraser Experimental Forest of central Colorado increase snow accumulations on the ground of treated areas (Meiman, 1987). While the responsible processes are not completely known, data on the effects of patch cuts and thinning treatments suggest that a reduction in snow interception loss is a major factor. Still inseparable, however, are the processes of interception and differential deposition of snow on a larger scale during snowfall events.

subalpine forest overstory were attributed to reduced interception of snow (Stednick and Troendle, 2004).

The process of interception during summer rainfall generally results in greater losses than with snowfall. The total interception loss is the sum of (1) water stored on vegetative surfaces (including plant residues – forest litter) at the end of a storm, and (2) the evaporation from these surfaces during the storm.

If a storm were to last over a long period under windy conditions, the interception loss would be expected to exceed that from a storm of equal duration under calm conditions. Conversely, a high-intensity and short-duration thunderstorm with high wind velocities can have the least amount of interception loss. Under such conditions water can mechanically be removed from the canopy and, therefore, not allow the storage capacity of the canopy to be reached. The effects of wind on evaporative loss would be minimal for a storm of short duration.

Potential interception loss (I) for a storm can be expressed as (Meriam, 1960 as presented by Gray, 1973):

$$I = S(1 - \mathrm{e}^{-P/S}) + RtE \tag{4.7}$$

where S is the water storage capacity of vegetative surfaces, expressed as depth over the projection area of canopy (mm); P is the rainfall (mm); e is the base of natural logarithms; R is the ratio of evaporating surface to the projected area (decimal fraction); t is the time duration of storm (h); and E is the evaporation rate (mm/h) during the storm.

The storage capacity of mature conifers is generally greater than that of mature hardwoods (deciduous trees). Comparisons among conifer stands have yielded a wide range of values in contrast to less variability among mature hardwood stands in North America. Interception losses of deciduous hardwoods vary with season as a result of leaf fall. The total interception loss in all forests can be attributed to the overstory tree canopy, understory herbaceous plants, and litter layer.

Hydrologic Importance of Interception

The hydrologic importance of interception is dependent on vegetative and precipitation characteristics and the climatic setting. We discuss interception in terms of its role in affecting annual water yield and watershed response to individual storm events. From a water budget viewpoint, the cumulative interception over time can be an important part of ET that represents a loss from measured *gross precipitation*. The result is *net precipitation* or that amount of precipitation available either to replenish soil water deficits or to become surface, subsurface, or groundwater flow. Net precipitation can be determined from

$$P_{\mathrm{n}} = P_{\mathrm{g}} - I \tag{4.8}$$

where P_{n} is the net precipitation (mm); P_{g} is the gross precipitation measured by rain gauges in openings (mm); and I is the interception loss (mm).

The above terms can be measured on individual plots (see Fig. 4.1) or estimated for designated forest stands using interception information from the literature. Determining net precipitation over a watershed is more difficult, however. The spatial variability of the tree canopy cover type and extent, canopy stratification (layering), and storage capacity of the litter all affect the total interception loss for a watershed. Under certain climatic conditions, the interception storage differences among species result in water yield differences. In regions where annual precipitation exceeds the *PET*, differences in interception between conifers and hardwoods can result in differences in water yield. For example, converting from hardwoods to conifers in the humid southeastern United States would be expected to increase interception losses and reduce annual water yield (streamflow and/or groundwater recharge). Such differences would likely not be observed in semi-arid regions because of the higher ratio of annual *PET* to annual precipitation. The difference in net precipitation reaching the soil surface due to differences in interception will satisfy soil water deficits rather than contribute to streamflow in dryland forests. The increase in net precipitation will simply be transpired at some later time and will not necessarily result in an increase in water yield.

Interception has been studied widely, and the literature cites interception values for many vegetative types. In excess of 20% to more than 40% of annual precipitation can be intercepted in forests while grassland ecosystems can intercept 10–20% of gross precipitation in periods when maximum growth has been attained. Regarding the latter situation, however, the plant residue on a tall grass prairie in Wisconsin has been reported to intercept 477 mm or 70% of the 681 mm of rain falling in a growing season (Brye et al., 2000).

The highest annual losses of water due to interception have been reported in coniferous forests in humid temperate regions and tropical forests (Table 4.2). Studies show that conifer forest interception generally exceeds that of deciduous hardwood forests and that forests in the temperate zones intercept a greater percentage of annual precipitation than tropical forests. Annual interception/precipitation relationships for 18 humid temperate conifer forests in the United Kingdom indicate that on average interception was approximately 40% of annual precipitation. Interception of redwood forests and old-age coniferous forests in the humid Pacific Northwest of the United States averaged almost 25% of annual precipitation.

TABLE 4.2. Interception of forest types

Forest type and location	Losses (mm)	% of annual P	Reference
Conifers – Wales	529/2013	26	Calder, 2005
Conifers – United Kingdom (18 locations)		25–48	Calder, 2005
Conifers – India		20–25	Ghosh et al., 1982 Chandra, 1985
Broadleaf – India		20–40	Ghosh et al., 1982 Chandra, 1985
Lowland tropical forests – West Java, Indonesia	595/2835 (P)	21	Calder, 2005
Wet tropical – Brazil	363/2593 (P)	14	Calder, 2005
Montane rainforests – Colombian Andes		12–18	Veneklaas and Van Ek, 1990
Pinyon-Juniper woodlands – Southwestern USA		5–10	Skau, 1964
Conifers – SW Washington USA		23–25	Link et al., 2004
Conifer–Redwood – California, USA		22	Reid and Lewis, 2009

TABLE 4.3. Interception storage for red pine stands at the Cloquet Forestry Center, Minnesota

Stand	Stand characteristics			Canopy storage		Litter storage capacity (in)
	Age (yr)	Basal area (ft^2/acre)	Stems (no./acre)	$P_g < 1$ (in)	$P_g > 1$ (in)	
A	21	85	1030	0.06	0.14	0.07
B	20	165	1512	0.14	0.28	0.12
C	29	234	1150	0.10	0.22	0.16
D	71	174	427	0.09	0.15	0.16

Source: Fox (1985).

Interception losses of forests in the humid tropics are more variable than in temperate regions. Many tree species in the humid tropics have large waxy leaves. This type of leaf and the frequent high-intensity rainfall favor smaller percentages of interception. Low-intensity, long-duration rainfall in coniferous forests of temperate climates with their high leaf surface area also tend to favor higher interception losses.

Annual interception losses in dryland forests are generally lower than those in either humid temperate or humid tropical forests because of lower canopy densities. However, in such climates, even small amounts of water loss can be important. For example, up to 70% of the late summer rainfall is intercepted by trees in oak-woodland communities of the southwestern United States and northern Mexico (Haworth and McPherson, 1991).

For any given forest type and climatic setting, the stand and litter characteristics need to be assessed with precipitation data to get a picture of the role of interception. Examples of red pine stand characteristics and associated interception storage values of the canopy and litter in northern Minnesota are presented in Table 4.3. These storage relationships were linked with measured precipitation data to calculate the growing-season interception losses for the respective stands (Table 4.4). The canopy interception storage represents both overstory and associated understory vegetation; litter interception storage is reported separately. Based on these tables, the interception loss for dense conifer plantations can be more than 30% of the precipitation occurring during the growing season.

The role of interception in reducing net rainfall during large storm events has generally been ignored by hydrologists in stormflow-flood studies. In some instances, however, interception by forests can be important. The amount, duration, intensity, and pattern of rainfall all influence the amount of interception. For example, the amount of interception in coniferous forests of the Pacific Northwest ranges from 100% for storm rainfall <1.5 mm to 15% for storm rainfall >75 mm (Rothacher, 1963). Rainfall interceptions losses were between 15% and 21% for high-intensity rainstorms up to 70 mm in a 120-year-old redwood forest in northern California (Reid and Lewis, 2009).

The hydrologic role of plant residues and forest floor litter interception is twofold: (1) the storage of part or all of the throughfall, and (2) the protection that litter provides for the mineral soil surface against the energy of rainfall. The storage capacity of forest litter depends on the type, thickness, and level of decomposition. In general, the storage capacity of conifer litter exceeds that of hardwood litter. Litter storage capacities of conifer

Table 4.4. Simulated interception components for the growing season (June, July, and August) for four red pine stands at Cloquet Forest Center, Minnesota, as described in Table 4.3

Year	Gross rainfall (P_g) (in)	Stand	Simulated net rainfall (in)	Canopy interception (in)	Litter interception (in)	Stemflow (in)
1953	21.55	A	18.19	3.50	0.42	0.57
		B	16.49	5.07	0.62	0.64
		C	17.46	4.02	0.62	0.45
		D	17.43	3.46	0.74	0.08
1970	6.13	A	4.68	1.16	0.41	0.12
		B	3.48	2.10	0.67	0.13
		C	3.87	1.66	0.69	0.09
		D	4.06	1.30	0.78	0.02
1976	11.74	A	9.79	1.77	0.46	0.28
		B	8.48	2.90	0.67	0.31
		C	9.02	2.23	0.71	0.22
		D	9.12	1.82	0.85	0.04

Source: Fox (1985).

plantations in Minnesota were similar to canopy storage capacities (see Table 4.3). However, the moisture content of litter usually remains high because the forest floor is protected from wind and direct solar radiation. As a result, litter might not be able to absorb much additional water, a point emphasized by contrasting litter storage capacities in Table 4.3 with seasonal litter interception losses in Table 4.4. Forest stands that are more open such as ponderosa pine in the southwestern United States can experience significant litter interception. Ponderosa pine litter storage capacities of more than 200% by weight have been reported (Clary and Ffolliott, 1969).

The protection against rainfall that litter provides for the soil surface influences the surface soil conditions directly and, therefore, infiltration, surface runoff, and surface soil erosion as discussed in Chapter 8.

Up to this point in our discussion, interception has been considered a loss from a watershed. However, along coastal areas and in high mountains near coastal regions, there are prolonged periods of low clouds or fog in which unique forest communities have developed. These are called *cloud forests* in which atmospheric moisture condenses on vegetative surfaces, coalesces, and drips from the canopy to add moisture to the soil that would otherwise remain in the atmosphere (see Chapter 12). This addition of moisture can exceed the interception loss of rainfall annually and result in a greater net rainfall under the canopy than gross rainfall in adjacent open areas. In such cases, the greater the foliage surface area, the greater the interception input to the water budget.

Transpiration

The flow of water through the soil–plant–atmosphere system is analogous to the flow of electrical current in an electrical circuit (Fig. 4.2a). The soil can be represented as a variable

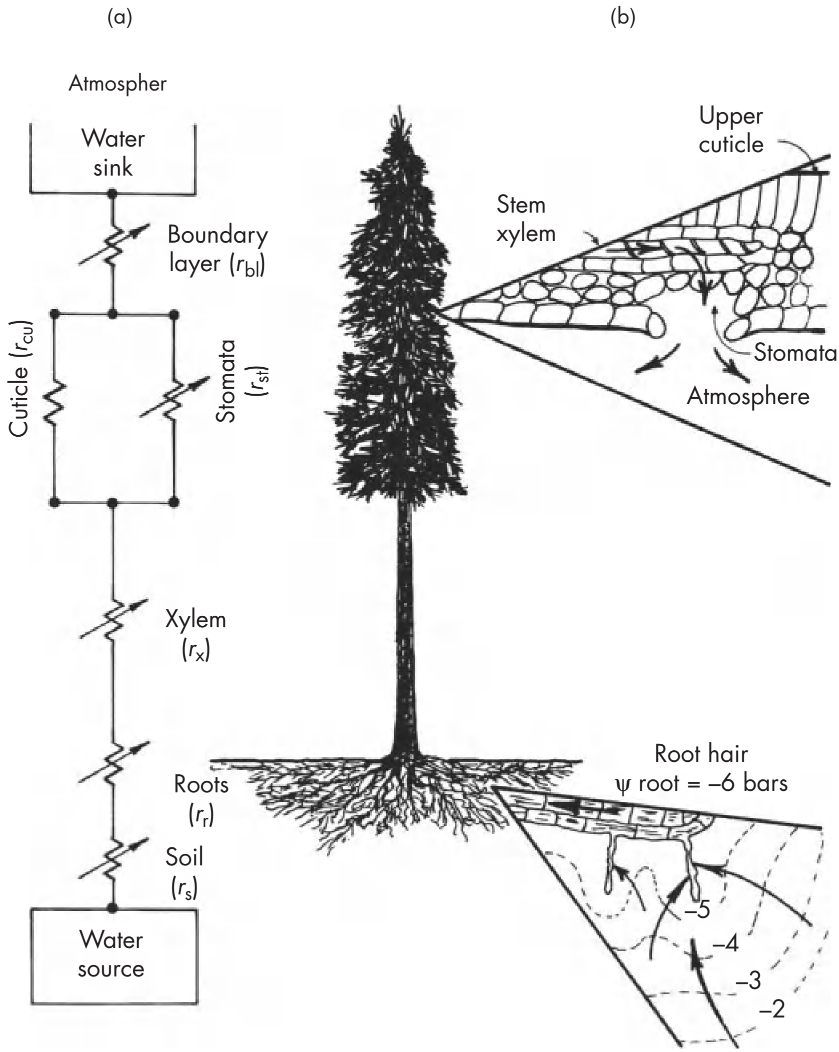


FIGURE 4.2. A representation of soil–plant–atmospheric resistances to water flow **(a)** and corresponding flow through the roots and leaves **(b)** (adapted from Rose, 1966, © Pergamon Books Ltd, with permission). The flow of water can be described by $V = \Delta\psi/r$, that is analogous to Ohm's law

resistor that changes with soil water content and soil–root interfaces. More resistance is offered to flow as the soil dries; however, the resistance to flow is offset when roots grow into moist soil.

Transpiration requires that liquid water must flow through the soil to plant roots (see Chapter 2) up the plant and to the *leaf–atmosphere* interface (Fig. 4.2b). Water flow through soils is relatively passive until intercepted by the roots of plants. Once the root absorbs soil water, different forces become operative as the major constituents of water potential and the water potential gradient that drives the flow of water into and through the plant. The primary components of plant water potential are:

$$\psi_{pl} = \psi_o + \psi_p + \psi_t + \psi_g + \psi_m \tag{4.9}$$

where ψ_{pl} is the plant water potent ial; ψ_o is the osmotic potential in the plant cells; ψ_p is the pressure potential (turgor pressure in the plant cells); ψ_t is the temperature potential; ψ_g is the gravity potential; and ψ_m is the matric potential.

Isothermal conditions are usually assumed within the plant system, thereby eliminating ψ_t. Matric potential, an important part of soil water potential, is a minor constituent of ψ_{pl} and is usually ignored. Gravitational potential is not generally considered for herbaceous plants but can be important in tall trees. For example, about 0.3 bar/m of tree height must be overcome for water to move to the top of a tree. Water potential gradients between cells of plants are, therefore, due to the interaction of the osmotic potential (ψ_o) with the pressure potential (ψ_p).

In metabolizing plant cells, fluctuations of solute concentrations affect the energy status of cellular water. When solutes are added to cellular water, the ψ_{pl} of the cell is lowered. This steepens the water potential gradient between surrounding cells, causing water to move through differentially permeable membranes into the cell. Water can enter from intercellular regions as well. The increased water content in the cell causes an increased turgidity that in turn opposes water entry. The final water potential of the cell is determined by these opposing forces, which can be expressed in terms of pressure as follows:

$$\psi_{pl} = P_t - P_o \tag{4.10}$$

where P_t is the turgor pressure and P_o is the osmotic pressure.

The analogy to Ohm's law can again be used in describing water flow through the plant:

$$q = -k_w A\left(\Delta\psi_{pl}\right) = \frac{-A\Delta\psi_{pl}}{r_{pl}} \tag{4.11}$$

where q is the water flow (cm^3/s); k_w is the water permeability (cm/s/kPa); A is the membranous area (cm^2); $\Delta\psi_{pl}$ is the water potential difference (kPa); and r_{pl} is the resistance of plant components (kPa $\times$ s/cm).

The leaf of a plant is the primary food-manufacturing center that maintains the water potential gradient by increasing solute concentrations by the process of photosynthesis. The osmotic potential gradient alone is probably sufficient to cause some water flow up the stem. Turgor pressure is maintained until stomata open at which time water escapes to the atmosphere and the pressure potential in the leaf diminishes, thereby steepening the total water potential gradient from root to leaf. In response to the gradient, liquid water moves through the cells of the plant to the leaf where evaporation occurs through open stomata. Water vapor then diffuses out through the leaf boundary layer to be dissipated by turbulent mass transport into the atmosphere.

Transpiration is a biological modification of the evaporation process because it is a function of the physiology and structure of the plant and the environment. This modification is more efficient than soil surface evaporation because of the large evaporating surface presented by plant foliage exposed to turbulent airflows above the soil boundary layer. Plants affect the amount of water transpired by stomatal regulation (the variable stomatal resistor in Fig. 4.2), structural and physiological adaptations, and rooting characteristics. In essence, plants provide a variable conduit for water to flow from the soil water reservoir to the active evaporating surface at the leaf–atmosphere interface. This conduit bypasses the higher resistance offered by dry surface soils. To understand the importance of transpiration and associated land management implications, the basic process is examined below.

Once liquid water reaches cell surfaces within the leaf, 585–590 cal/g for temperatures of most terrestrial systems are required for vaporization. After vaporization, the water vapor flows through intercellular spaces to the substomatal cavity between guard cells of stomata and into the atmosphere in response to the *vapor pressure gradient* at the leaf surface. Water vapor can also take a parallel path through leaf cuticles. However, this pathway usually offers more resistance r_{cu} to flow than through stomata except when stomata are closed tightly. As a consequence, r_{cu} is considered large, making the variable resistor of stomata (r_{st}) the primary regulator of transpiration. The magnitude of r_{st} is proportional to the degree of opening of the stomatal pore, or stomatal aperture while the magnitude of r_{cu} is a function of cuticular integrity and thickness.

The total resistance offered to vapor flow by the leaf (r_l) is

$$\frac{1}{r_l} = \frac{1}{r_{st}} + \frac{1}{r_{cu}} \tag{4.12}$$

or

$$r_l = \frac{r_{st} r_{cu}}{r_{st} + r_{cu}} \tag{4.13}$$

The effect that the stomatal opening has on transpiration depends on the thickness of the boundary layer (r_{bl}) surrounding leaf surfaces; this is evident in Figure 4.2 because r_{bl} is in series with r_{st}. The total diffusive resistance (r) is described by

$$r = r_{bl} + r_l \tag{4.14}$$

Because boundary layer resistance is inversely related to wind speed, r_{bl} will be large under still-air conditions, causing r_{st} to have less effect on transpiration. Under windy conditions, changes in stomatal aperture strongly affect rates of transpiration (Fig. 4.3).

The vapor pressure gradient is necessary for vapor to flow from the substomatal cavity and is maintained by energy inputs to the leaf. This energy causes the vapor pressure of leaf water to be greater than the partial pressure of water vapor in the surrounding atmosphere. Therefore, more water molecules exit the liquid–air interface than enter and the gradient in liquid-water potential, $d\psi/dx$, is steepened from the leaf down to the root surface. Water flows into the plant in response to $d\psi/dx$ until the soil becomes so dry that the permanent wilting point is reached (Fig. 4.4).

Measurement of Transpiration

Most methods of measuring transpiration have been developed for agricultural crop plants, few of which are applicable for field measurements of trees or shrubs. For the most part, transpiration cannot be measured directly in the field without some type of disturbance to the plant. This section of the chapter provides a brief overview of some of the most common methods used in the field.

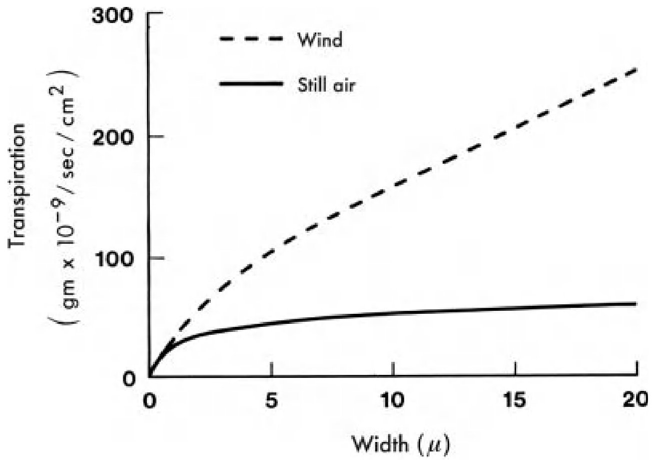


FIGURE 4.3. Relation between stomatal width and transpiration in still-air and windy conditions (from Slatyer, 1967 after Bange, 1953, with permission)

Potted Plants and Lysimeters

Soil–plant systems contained in small covered pots or larger tanks, called *lysimeters*, can be used to measure transpiration. Water budget analyses, in which every component is measured directly except transpiration, are performed on these systems.

The potted-plant method is suited for small individual plants. The bottom of the pot is perforated to allow water to drain freely. Typically, the soil is wetted thoroughly, the soil surface is covered (usually with plastic sheeting) to prevent soil evaporation, and the soil is allowed to drain. After all free water drains from the soil, the potted plant is weighed;

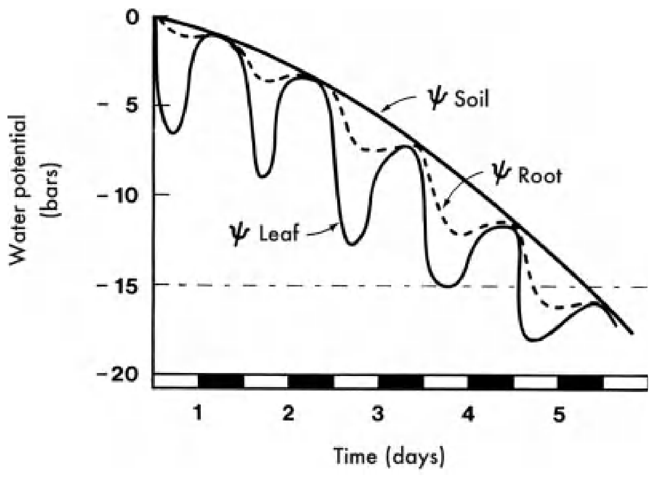


FIGURE 4.4. Changes in water potential (ψ) of soil, root, and leaf as transpiration occurs beginning with a soil near field capacity and proceeding until the permanent wilting point (–15 bars) is reached (from Slatyer, 1967, with permission)

the soil is assumed to be near *field capacity* at this point. At some determined time, the potted plant is reweighed, and the difference in weights is equated to transpiration loss over the period. This method is not suited for large plants, but small trees and shrubs can be measured and transpiration rates can be compared for different environmental conditions or treatments.

Lysimeters are tanks designed to hold a larger mass of soil and usually more than one plant and can be either a weighing type (similar to the potted-plant method) or a drainage type. With the drainage type, any surface runoff or drainage from the bottom is collected and measured. The only unmeasured part of the water budget is transpiration or total *ET*. The soil surface must be covered to obtain estimates of transpiration. Lysimeters are more commonly used to measure total *ET* for agricultural crops.

Because of the greater volume of soil, lysimeters allow plant roots to develop more naturally, and boundary conditions are not as severe as with potted plants. Although usually designed for smaller plants, Fritschen et al. (1977) developed a lysimeter to measure transpiration of a 28-m high Douglas-fir tree. The lysimeter weighed 28 900 kg and could detect changes in weight of 6.3 g. The difficulty of construction and costs associated with such lysimeters make them impractical for most situations.

Measurement of Sap Flow in Woody Plants

Transpiration rates for trees and shrubs can be estimated by measuring the net rate of the accent of xylem sap through stems. This method allows transpiration to be examined across different sites and stand structures (Ford et al., 2007). The fundamental basis for this method is that most of the water that moves up the stem of a plant is largely in response to transpiration. Other causes of sap flow can be ignored since there is a negligible amount of change in stem storage over time and only 2–5% of sap flow is used in photosynthesis. Two basic sap flow methods are the heat-pulse or tracing method, and the steady state or stem heat-balance method.

The heat-pulse or tracing method measures the velocity of a short pulse of heat, or a tracer such as deuterium (a heavy hydrogen isotope), moving up the xylem sap of woody stems. Sensors are located in the xylem both upstream and downstream of the source of heat or other tracer, and measurements are taken over time to determine net upstream flow of sap (Fig. 4.5). Sap flow velocity is useful for detecting when and at what rates sap flows in response to transpiration. Hinckley et al. (1978) reported maximum sap velocities ranging from 1550 to 6000 cm/h for ring porous species such as oak (*Quercus*) and 75 to 200 cm/h for nonporous coniferous species. To estimate the volume of water loss per plant, the sap flow velocity needs to be related to the area of the stem through which sap flows.

The steady state or stem heat-balance method directs a known amount of heat into the stem and uses sensors that are external to the stem. The method is a heat-balance approach that relates heated stem tissue to sap flow rates. When water flow through the stem is relatively constant, the temperature of the sapwood will reach a stable value (Swanson, 1994). If there is minimal heat loss, the heat that is transported out of a section of stem is related to the amount of heat added to the section. Methods using constant heating relate sap flow to heat balance, thermal dissipation or heat field deformation (Cermak et al., 2004; Lu et al., 2004).

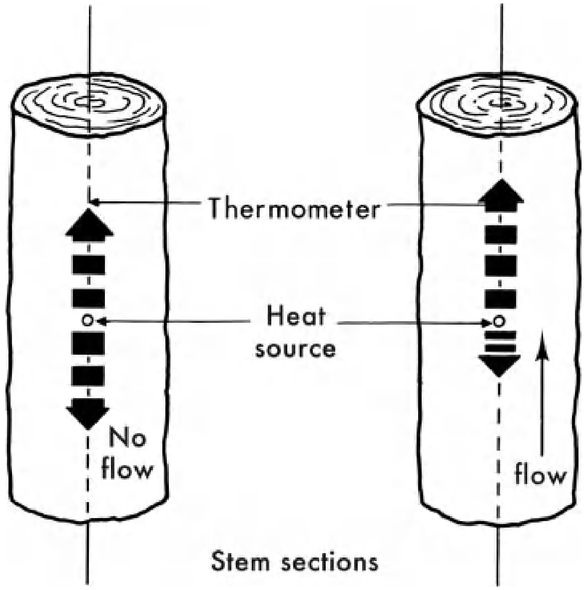


FIGURE 4.5. Sap flow velocity method using heat as a tracer; heat dissipation in stem on the left indicates no transpiration while the stem on the right indicates transpiration (from Swanson and Lee, 1966)

An application of the sap flow method for determining annual transpiration rates of oak trees is presented in Box 4.3.

Tent Method

In the tent method, a plant is enclosed with plastic sheeting, and the rate and the moisture content of air entering and leaving the tent are monitored (Fig. 4.6). When the amount of moisture in the air leaving the tent exceeds that entering the tent, the difference is due to transpiration if the soil surface is covered or *ET* if soil evaporation is permitted. The method excludes rainfall interception from the *ET* process.

This method facilitates measurement of the transpiration of large shrubs and small trees without transplanting or disturbing the soil. One disadvantage is the buildup of heat in the tent caused by the trapping of radiation, that is, the greenhouse effect. This problem can be partly overcome by maintaining adequate air circulation in the tent. The same principle is used for small enclosures for individual leaves. Nevertheless, the environment surrounding the plant is artificial, and the measured transpiration rates might not coincide with rates outside a tent. The method can be used to indicate relative transpiration differences between plant species in adjacent plots or the same species under different treatments.

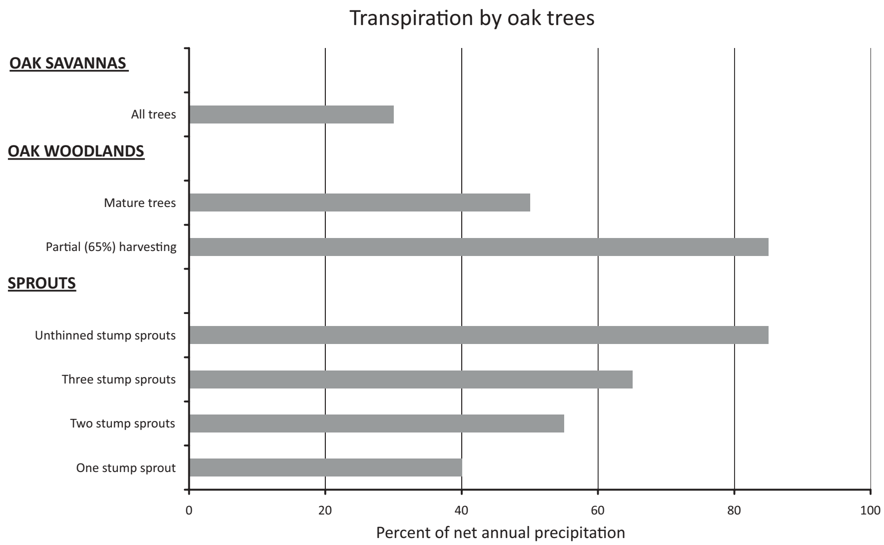
Other Methods

There are several other methods of estimating transpiration activity that do not indicate rates or volumes of whole plant transpiration over time. One such method, called *quick weighing*, has been used to overcome the difficulty of weighing large plants in the field. A

Box 4.3

Annual Transpiration of Oak Trees in Southeastern Arizona (Ffolliott et al., 2008)

Annual transpiration of oak trees obtained by sap flow measurements have been expressed as a percent of annual precipitation (450 mm) for varying stand conditions and silvicultural treatments in southeastern Arizona. Presented in the following figure is annual transpiration of trees on sites in a low-density oak savanna and a higher density oak woodland. The two sites were similar in the past and present land-use practices imposed on them. Values for a mature (uncut) stand and a stand in which 65% of the trees in the oak woodlands were harvested for firewood are also shown. The increased transpiration in the harvested stand is attributed to the large number of post-harvesting stump sprouts in the site. While the annual transpiration of an individual stump sprout was comparatively small, the large number of stump sprouts in the harvested stand translated into the larger transpiration value.



Annual transpiration of the stump sprouts on rootstocks of trees in the stand harvested for firewood after a thinning treatment is also compared to annual transpiration on a site where the stump sprouts were not thinned. Note that the annual transpiration of stump sprouts on the unthinned site is similar to the transpiration of stump sprouts on partially harvested site because of the overwhelming dominance of stump sprouts.

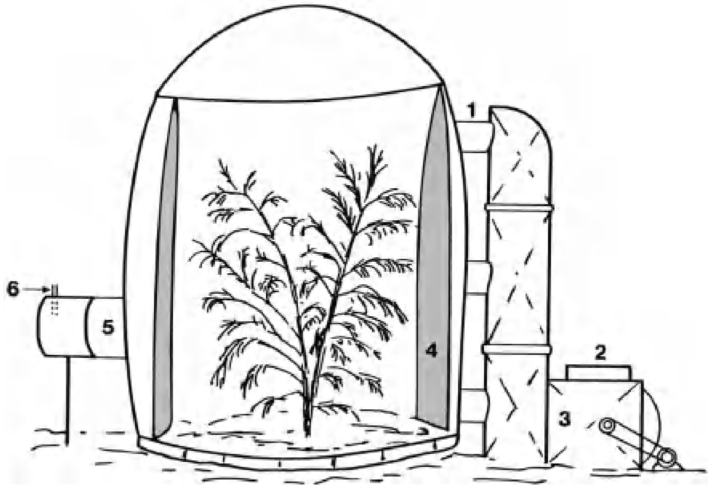


FIGURE 4.6. Triple-inlet evaporation tent (from Mace and Thompson, 1969). 1, inlet; 2, squirrel-cage blower; 3, inlet humidity thermometer; 4, perforated polyvinyl curtain; 5, outlet; and 6, outlet humidity thermometer

leaf or small branch is cut off, weighed immediately, then reweighed after a short period. The change in weight is related to transpiration. The severity of plant disturbance makes this a questionable method. However, one can obtain comparative data in the field that can be used to indicate relative transpiration activity.

Porometry is another indirect transpiration measurement method that temporarily encloses transpiring leaves or shoots in a small chamber within which changes in humidity are measured (Shuttleworth, 2008). This method allows estimates of transpiration due to stomatal responses to different environmental influences, such as orientation of leaves on a tree within stand, at different heights, and so forth. Extrapolating transpiration of individual leaves to whole plants would require considerable sampling of leaves and usually would not be practical for trees.

Transpiration and Interception Relationships

When a vegetative surface intercepts rainfall, part of the energy normally allocated for transpiration is used in evaporation at the leaf surface. Some compensation occurs in that transpiration rates are often reduced when the foliage is wet. The net effect, however, is usually a greater total loss of water by vaporization than would have occurred via transpiration alone. Evaporation rates of wet canopies have been reported to be two to three times greater than transpiration rates for forest stands largely because evaporation of a wet leaf surface is not affected by stomatal resistance. Also, a greater exchange of advective energy results in high evaporation rates because forest canopies are rough surfaces projected into the more turbulent upper air. Wet forest canopies evaporate water at higher rates than wet low-growing agricultural crops or grasses and can exceed *PET* rates as estimated by traditional methods discussed later in this chapter. Furthermore, evaporation rates at night can far exceed transpiration rates because the stomata of most plants close at night.

Effects of Vegetative Cover

The type, density, and coverage of plants influence transpiration losses. Differences in transpiration rates among individual plants and plant communities can be attributed largely to differences in rooting characteristics, stomatal response, and albedo of plant surfaces. Annual transpiration losses are also affected by the duration of a plant's growing season. Herbaceous vegetation such as grasses and agricultural crops generally has shorter growing seasons and, as a result, shorter active transpiration seasons than woody shrubs and forest vegetation. Likewise, deciduous forests normally transpire in a shorter season than conifers. Transpiration rates of trees are usually higher than most other plants. Wullschleger et al. (1998) reviewed 52 studies conducted since 1970 and reported whole tree water use for 67 species in over 35 genera. Some of these maximum daily rates of transpiration are presented in Table 4.5.

Comparing a bare soil, a herbaceous grassland, and a mature forest can illustrate the effects of changing vegetative cover on transpiration and total *ET* (Fig. 4.7). If soil water is abundant in all three sites, evaporation and transpiration will occur at rates primarily dependent on available net energy, vapor pressure gradients, and wind conditions. Differences in overall vapor loss will be largely the result of differences in advective energy. Transpiration by plants with a large leaf area and a canopy extending higher above ground can surpass that of smaller plants in such instances. With a dense forest canopy, advection might only affect transpiration of trees at the edges of stands or forest openings.

TABLE 4.5. Maximum daily transpiration rates for selected tree species

	Method	Height (m)	Diameter (cm)	Leaf area (m2)	Water use (kg/day)
Conifers					
Abies amabilis	TM	18	40	151	98
Larix gmelinii	TM	20	25		67
Picea abies	TM	25	36	447	175
	TM	17	15		66
Pinus caribaea	R/SI	7	13		100
Pinus contorta	L/P		20–26		44
Pinus radiate	TM	25	42	300	349
Pseudotsuga menziesii	L/P	28	38		64
	R/SI	76	134		530
Hardwoods					
Eucalyptus grandis	L/P	6	41	219	174
	TM	34	30	71	141
Eucalyptus regnans	TM	58	89	330	285
Populus x euramericana	L/P	5	—	26	109
Populus trichocarpa x P. deltoides	TM	15	15		51
Quercus petraea	TM	15	9		10
Salix matsudana	L/P	5		28	106
Spondias mombin	TM	23	44		80

Source: Wullschleger et al. (1998).

TM, thermal balance or heat dissipation (sap velocity); R/SI, radioactive or stable isotopes (sap velocity); L/P, lysimeters or large tree photometers.

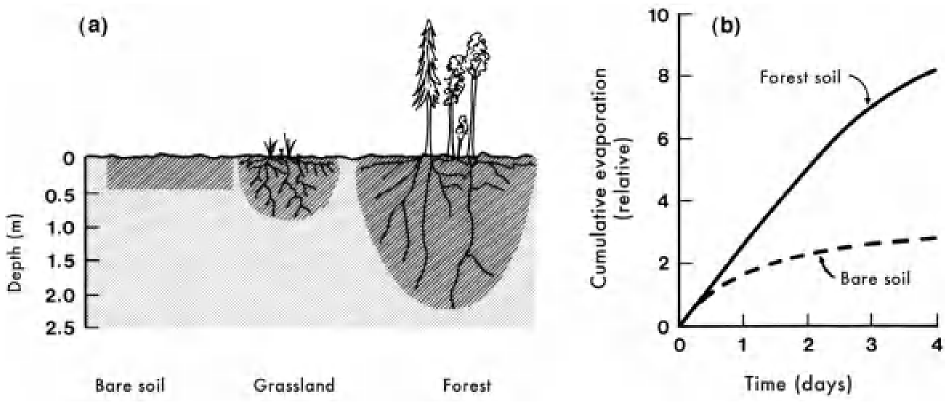


FIGURE 4.7. Effects of changing vegetative cover on transpiration and total *ET*. **(a)** Soil water depletion (cross-hatched areas) of a bare soil, a grassland, and a mature forest (vertical scale exaggerated). **(b)** The associated accumulated evaporation–evapotranspiration for bare soil conditions (from Lee, 1980 © Columbia University Press, with permission)

Plant canopies can also increase the rate of vapor flow by creating more turbulent airflow around transpiring surfaces. This effect would be particularly pronounced with some conifers, whose needlelike leaves create numerous small eddies. Therefore, larger-canopied trees can potentially transpire larger amounts of water than would otherwise evaporate from a bare soil or transpire from communities dominated by herbaceous plants of smaller stature such as grasses and forbs.

Once the soil begins to dry, the slower rate of water movement through the drier soil limits the rate of evaporation at the soil surface. Soil water depletion can occur only to a depth of 0.4 m after a given period. Except for very coarse-textured soils, evaporation seldom depletes soil water below a depth of 0.6 m. The flow of water to the evaporating–transpiring surface of herbaceous vegetation can continue for a longer time because plant roots grow and extend into greater depths (1 m in this example of a grassland, see Fig. 4.7a) and extract water that would otherwise not evaporate from a bare soil in the given time.

Deep-rooted forest vegetation is capable of extracting water to depths greater than 2 m and, therefore, has greater access to the soil water reservoir. Over time, the differences in evaporation from a bare soil versus *ET* from a forest can be substantial (see Fig. 4.7b). Such differences in soil water depletion can result in differences in water yield. For a given rainfall or snowmelt event, more water is required to recharge soils under forest vegetation than soils with herbaceous cover. The least amount of water would be needed to recharge bare soil areas. As a consequence, the proportion of rainfall or snowmelt that becomes streamflow will generally be greatest for the bare soil and least for the forested area.

The preceding discussion applies to watersheds on which soil moisture controls *ET*, which would not necessarily be the case for wet climates where rainfall is prevalent throughout the growing season. Also, different species of plants exert variable control of *ET* through stomatal responses that can be linked to different climatic factors such as wind.

TABLE 4.6. Controls of ET under different climatic conditions and types of vegetation

Type of vegetation	Dry climate	Wet climate
Temperate climate		
Forests – tall vegetation	Physiological Soil water	Advection
Annual crops – short vegetation	Soil water Physiological	Radiation Physiological
Tropical climate		
Forests – tall vegetation	Soil water Tree size	Drop size Physiological
Annual crops – short vegetation	Soil water	Radiation

Source: Adapted from Calder (1998, 2005), with permission from Oxford University Press.

A synthesis of studies in temperate and tropical regions of the world that helps us understand *ET* responses to changing vegetation and land use under different climatic regimes is presented in Table 4.6. Factors controlling *ET* have been found to differ between tall vegetation (e.g., forests) and short vegetation (e.g., annual agricultural crops) in wet and dry climates. Soil water content is a controlling factor with all types of vegetation in dry climates. However, advective energy is the controlling factor that is related to the high levels of interception losses from forests in the wet climates of temperate zones, as discussed earlier. *ET* from short vegetation in the same climatic regime is more dependent on radiation energy and physiological controls. High interception losses from forests in the wet tropics are related to rainfall-drop size because of the large waxy leaves of most of the tree canopies. Transpiration rates of trees in the wet tropics are also governed more by physiological responses. Radiation appears to control *ET* of annual crops in the wet tropics.

The ability to predict changes in water yield depends on having methods or models capable of predicting *ET* in different climates. Many such methods rely on understanding the potential evapotranspiration (*PET*) that exists in a particular location and relating the actual *ET* to that *PET*.

POTENTIAL EVAPOTRANSPIRATION

The concept of *PET* has its origin in *ET* studies of irrigated crops. *PET* was defined by Penman (1948) as the amount of water transpired in a unit of time by a short green crop, completely shading the ground, of uniform height, and never short of water. This definition was supposedly an expression of the maximum *ET* that could occur and was limited only by available energy or radiation in the case of a short plant crop. This led to thinking that all well-watered soil–plant systems and open bodies of water will lose equal amounts of water with these amounts controlled by available energy. Available energy (discussed earlier) can differ appreciably among vegetative types, different soils, and bodies of water. Empirical models of energy availability have been developed as workable definitions or indices of *PET*.

Several methods of estimating *PET* are described in the literature but only a few will be reviewed here. In all cases, equations and relationships used to approximate *PET* should be considered only as indices of *PET*. Before applying any *PET* equation one should review its origin and understand its limitations and range of applications.

Empirical relationships can be used to approximate *PET*. Those relationships that require only air temperature data are attractive to hydrologists because they are simple and the data are readily available in most areas. Thornthwaite's equation (Thornthwaite and Mather, 1955) estimates *PET* for 12-hour days and a 30-day month with the following equation:

$$PET = 1.6\left[\frac{10T_{\mathrm{a}}}{I}\right]^{a} \tag{4.15}$$

where *PET* is the annual potential evapotranspiration (cm); T_{a} is the mean monthly air temperature (°C); and I is the annual heat index, which can be given by

$$\sum_{i=1}^{12}\left(\frac{T_{\mathrm{a}i}}{5}\right)^{1.5} \tag{4.16}$$

and $a = 0.49 + 0.0179\,I - 0.0000771\,I^2 + 0.000000675\,I^3$.

Values of *PET* determined thus must be adjusted for the number of days per month and day length (latitudinal adjustment).

Hamon's equation (1961) estimates *PET* by

$$PET = 5.0\ \rho_{\mathrm{s}} \tag{4.17}$$

where *PET* is the daytime potential evapotranspiration (mm/30-day month) and ρ_{s} is the saturation vapor density at mean air temperature (g/m^3).

Penman (1948) combined a simplified energy budget with aerodynamic considerations to estimate evaporation. The Penman equation, a widely known method of estimating daily *PET*, is defined as

$$PET = \frac{\Delta R_{\mathrm{n}} + \gamma E_{\mathrm{a}}}{(\Delta + \gamma)L} \tag{4.18}$$

where *PET* is the evaporation or potential evapotranspiration (g/cm^2/day); Δ is the slope of saturation vapor curve (mb/°C); γ is the psychometric constant (0.66 mb/°C); R_{n} is the net radiation (cal/cm/day); $E_{\mathrm{a}} = (e_{\mathrm{s}} - e_{\mathrm{a}})\,f(u)$, where $e_{\mathrm{s}} - e_{\mathrm{a}}$ is vapor pressure deficit at 2 m height (mb), $f(u)$ is the wind function (km/day), approximating atmospheric diffusivity near evaporating surface (g/cm^2/day); and L is the latent heat of vaporization (585 cal/g at 21.8°C).

Penman's equation has been modified to include plant coefficients that express physiological and aerodynamic resistance of vegetation. Modifications were based on evaporation and transpiration experiments of forest vegetation that exposed the weaknesses of *PET* methods (Calder, 1982, 2005; Morton, 1990). Early work implied that all wet vegetative surfaces experienced largely similar *ET* with any differences attributed to albedo. This assumption was attractive to practitioners because maximum possible rates of *ET* could be estimated from physical and meteorological measurements. However, experimental evidence indicates that:

- *ET* losses vary significantly among vegetative types (e.g., forests vs. grasses) even if both systems have abundant available water.
- Forest interception losses can exceed *PET* as predicted by Penman's equation.

- Some forest stands that are “well supplied” with water will periodically transpire significantly less water than would be predicted from net radiation methods because of physiological controls.

The departures from earlier *PET* concepts addressed above can be explained by the Penman–Monteith equation (Monteith, 1965):

$$PET = \frac{\Delta R_n + \rho C_p (e_s - e_a)/r_a}{\Delta + \gamma (1 + r_{st}/r_a)} \tag{4.19}$$

where ρ is the air density; C_p is the specific heat of air; r_a is the aerodynamic resistance; and r_{st} is the plant stomatal resistance.

When a film of water covers leaves (interception), r_{st} becomes negligible. In addition, r_a for forest vegetation is smaller than that for grasses and other low-growing herbaceous vegetation. As much as 80% of the total energy input to wet forest canopies can be derived from advection; even in humid climates such as that in England, latent heat flux can exceed net radiation by 12%.

Equation 4.19 also helps to explain the differences in transpiration losses among species that cannot be explained based on energy exchange alone. Plants control transpiration through stomatal response. Stomata respond to changes in light intensity, soil moisture, temperature, and vapor pressure deficit. Most plant species close their stomata at night. Some close their stomata in response to wind. Many species close stomata when soils become dry.

ESTIMATING ACTUAL EVAPOTRANSPIRATION

At the watershed scale, *ET* losses are affected by net radiation, advection, turbulent transport, and vegetative characteristics including plant structure, leaf area, stomatal resistance, canopy resistance, and plant water availability (Zhang et al., 2001). We already have seen that when climatic conditions differ, the factors that have major control over *ET* of forest or short agricultural crops differ as well (see Table 4.6). As a result of the variable factors that affect actual *ET*, there are no practical field methods available to measure *ET* on a watershed landscape.

The fast-moving developments of new technologies and improved instrumentation in the early twenty-first century suggest that improved methods of measuring *ET* might be forthcoming. Measurements of *ET* at watershed scales include remote sensing estimates and applications of Light Detection And Ranging (LIDAR) method (Shuttleworth, 2008). Problems with non-uniform aerodynamic roughness and surface temperature–air temperature relationships and atmospheric stability affect one or more of such methods. Readers can use internet search engines to learn about the latest developments and applications of these technologies.

Local estimates of *ET* for a particular vegetative type or uniform land cover conditions can be made with energy budget micrometeorological approaches such as the Bowen Ratio and Eddy correlation (eddy covariance). The Bowen ratio (β) uses the ratio of sensible to latent heat derived from atmospheric temperature (ΔT) and humidity (ΔE_o) gradients (Shuttleworth 1993, 2008):

$$\beta = \gamma \left(\frac{\Delta T}{\Delta E_o} \right) \tag{4.20}$$

where γ is a constant to account for units.

By measuring net radiation R_n and soil heat flux (G), the evaporation rate (E) in mm/day can be estimated as

$$E = \frac{R_n - G}{1 + \gamma\,(\Delta T/\Delta E_o)} \quad \textbf{(4.21)}$$

The Bowen ratio method provides reasonable estimates for field estimates of short crops but is problematic for tall vegetation such as that found in forests.

The Eddy correlation method uses time increment averages from the correlation coefficient between vertical fluctuations in wind speed and atmospheric humidity measured at high frequency a few meters above the ground surface (Shuttleworth, 2008). Instrumentation is complex and expensive. A reader interested in greater detail should access the internet on the latest technology and instrumentation.

For most watershed-level studies, the best estimates of vegetative effects on actual *ET* come from a combination of process studies, water budget analyses, and paired-watershed experiments (Box 4.4). An example of the latter was an experiment conducted at Hubbard Brook, New Hampshire, in which two similar watersheds with mixed hardwood forests were calibrated and one was subsequently cleared of all living vegetation (Hornbeck et al., 1970). By suppressing vegetative regrowth with herbicides on the cleared watershed, water yield was increased an average of 290 mm/ year over a 3-year period; these changes were largely due to reduced transpiration. An accurate determination of total transpiration was not possible because evaporation from soil and litter was not suppressed.

The absence of practical methods of measuring *ET* requires that hydrologists and watershed managers apply their knowledge of *ET* processes of plant and soil systems with which they are working. A commonly used approach is to employ some index of *PET* and relate *PET* to available water in the watershed. Such an approach requires knowledge of soil water characteristics and plant response to soil water changes.

Evapotranspiration/Potential Evapotranspiration Approach

Relationships between transpiration or *ET* and soil moisture deficits (soil water content below field capacity) have been developed for a few wildland species. Relationships of the form below have been used to relate actual *ET* to *PET*:

$$ET = (PET)\, f\,(AW/AWC) \quad \textbf{(4.22)}$$

where f is the functional relationship; *AW* is the available soil water (mm) = (soil moisture content − permanent wilting point) × rooting depth of mature vegetation; and *AWC* is the available water capacity of the soil (mm) = (field capacity − permanent wilting point) × (rooting depth of mature vegetation). Field capacity (*FC*) for different soil textures can be estimated (see Chapter 2).

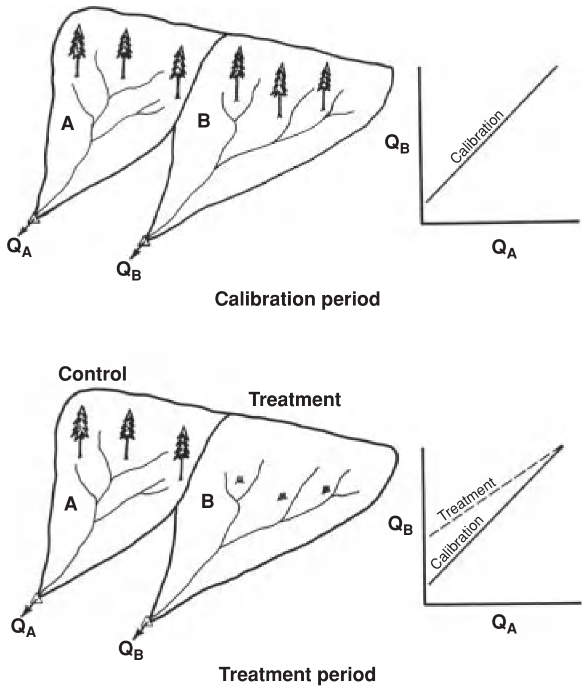
For most temperate forests where soil moisture controls *ET*, the actual *ET* would normally be expected to be at or near *PET* when soil moisture conditions are near field capacity. However, quantifying the relationship $f(AW/AWC)$ requires experimental evidence of plant or plant community response to soil moisture deficits. Tan and Black (1976) found that transpiration rates of Douglas-fir trees were halved when ψ_s equals −1000 kPa. These authors also indicated that high vapor pressure deficits accentuated the effects of soil moisture deficits.

Box 4.4

The Paired-Watershed Method as a Means of Estimating the Effects of Timber Harvesting on Streamflow

This method is often used for evaluating the effects of timber harvesting on streamflow. Two watersheds first must be selected on the basis of similar soils, vegetative cover, geology, and topography. They should be close to one another and of similar size. Streamflow is measured at the outlet of both watersheds (as discussed in Chapter 6) over a sufficient time so that streamflow from one (watershed B in the figure below) can be predicted from streamflow measurements at the other (watershed A in the figure below). Depending on the similarity of the two watersheds, usually from 5 to 10 years is needed to achieve an acceptable regression relationship for annual streamflow; this time period is called the *calibration period*. After calibration, the vegetation on watershed B can be cleared; however, streamflow measurements continue for both watersheds. Observed streamflows Q_B after treatment are then compared to streamflow values from watershed B that were predicted from the regression relationships based on streamflow measurements at Q_A. The treatment period following vegetation clearing shows an increase in streamflow at watershed B. The change in streamflow is largely attributed to changes in *ET*.

Paired watershed experiment



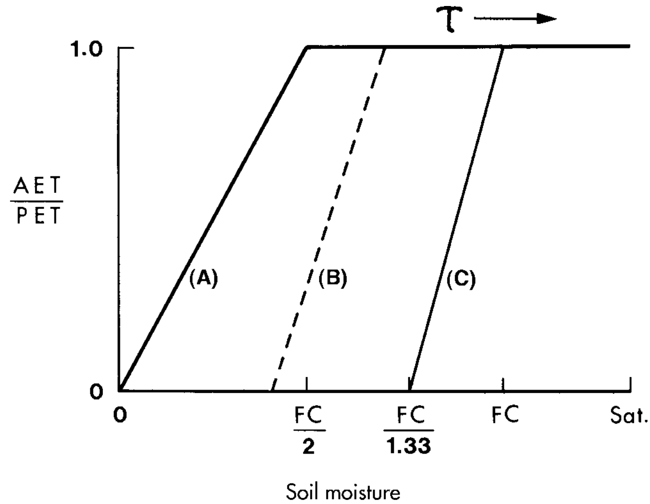


FIGURE 4.8. Actual evapotranspiration (*AET*) to potential evapotranspiration (*PET*) ratio as a function of water conditions for an old-growth forest **(A)**, intermediate forest cover condition **(B)**, and open or clearcut condition **(C)** (adapted from Leaf and Brink, 1975). FC, field capacity

Leaf and Brink (1975) developed relationships for forest types and stand conditions for coniferous forests in the Rocky Mountains of the United States. Among these relationships, the effects of clearcutting and regrowth on *f(AW/AWC)* are illustrated in Figure 4.8. Remember that *PET* values are only an index in all of these approaches.

The point at which soil water deficit begins to limit *ET* varies with stand condition from clearcut to old-growth trees (Fig. 4.9) and can be estimated from:

$$\tau = (FC)\,e^{-k(t-t_c)} \tag{4.23}$$

for conditions $\tau = FC$, $t < t_c$, and $\tau = FC/2$, $t > t_r$, where τ is the soil water deficit value at which *ET* becomes limited; k is the index of rate of decline of τ; t_c is the time in years when soil water begins to limit *ET*; and t_r is the time in years when the hydrologic effect of clearcutting becomes negligible.

The approach used in the above example requires information that might not be available for many types of vegetation. Simpler approaches can be used to approximate *ET* on a watershed basis.

Water Budget Approach

The concept of water budget as a hydrologic tool is relatively simple. If all but one component of a system can be either measured or estimated, then the unknown component can be solved directly. As described in Chapter 2, annual *ET* can be approximated from $P - Q$ as the annual change in soil moisture on an annual basis is usually negligible.

Applying the water budget method to estimate *ET* requires that all outflow of liquid water from the watershed has been measured and (sometimes) assuming that there was no loss of water by deep seepage to underground strata and that all groundwater flow from the

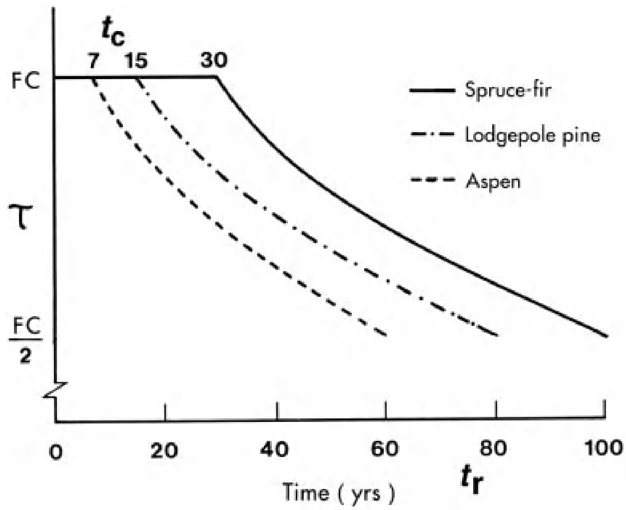


FIGURE 4.9. Effects of time and tree species on soil water deficit (adapted from Leaf and Brink, 1975). τ, limiting soil water deficit with time for the tree species; t_c, time (in years) when soil water begins to limit *ET*; t_r, time (in years) when the hydrologic effect of clearcutting becomes negligible

watershed was measured at the gauging site. If geologic strata such as limestone underlie a watershed, the surface watershed boundaries might not coincide with boundaries governing groundwater flow. In such cases, there are two unknowns in the water budget – *ET* and groundwater seepage.

If groundwater exchanges are negligible, the water budget approach can be used for purposes other than estimating actual *ET*. For the example in Box 4.5, actual *ET* is assumed to equal *PET* if soil moisture content is sufficient. Once available soil moisture is less than *PET*, the actual *ET* equals the available soil moisture content. This approach likely overestimates *ET* but it is useful for providing conservative estimates of water yield (runoff in Box 4.5) from a watershed. If *ET*/*PET* relationships are known (such as in Fig. 4.8), they should be used in the analysis.

SUMMARY AND LEARNING POINTS

The importance of *ET* and its influence on the water budget of a watershed should be recognized after reading this chapter. *ET* is one of the most significant hydrologic processes affected by human activities that alter the type and extent of vegetative cover on a watershed. Anyone on a watershed who manipulates soil–plant systems on the watershed should have an understanding of the process of *ET* and the factors that influence its magnitude. After reading this chapter, you should be able to:

(1) Explain and differentiate among the processes of evaporation from a water body, evaporation from a soil, interception loses from plants, and transpiration from plants.
(2) Understand and be able to solve *ET* using a water budget and an energy budget method.

Box 4.5

Water Budget Exercise for a Forest-Covered Watershed, Near Chiang Mai, Thailand

	Year 1				
	Oct	Nov	Dec	Jan	Feb
			(mm)		
1. Average rainfall[a]	130	46	10	5	10
2. Initial soil moisture[b]	192	192	131	39	0
3. Total available moisture	322	238	141	44	10
4. Potential ET[c]	114	107	102	99	85
5. Actual ET[d]	114	107	102	44	10
6. Remaining available moisture	208	131	39	0	0
7. Final soil moisture[e]	192	131	39	0	0
8. Runoff[f]	16	0	0	0	0

[a] Average over the watershed for each month of record.
[b] At start of each month. Same as "final soil moisture" of previous month.
[c] Average evapotranspiration (*ET*) annual values for the month, as estimated by Thornthwaite's method.
[d] Total available moisture, or potential *ET*, whichever is smaller.
[e] At end of month. Same as "initial soil moisture" for next month. This value cannot be larger than the soil-water-holding capacity determined for the watershed, for this watershed 192 mm; effective rooting depth = 1.2 m × 160 mm/m of available water (field capacity – permanent wilting point).
[f] Runoff occurs when the remaining available moisture exceeds the water-holding capacity for the watershed (192 mm).

(3) Explain *PET* and actual *ET* relationships in the field. Under what conditions are they similar? Under what conditions are they different?
(4) Understand and explain how changes in vegetative cover affect *ET* under different climatic conditions.
(5) Describe methods used in estimating transpiration of plants, *PET*, and actual *ET*.

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CHAPTER 5 *Infiltration, Pathways of Water Flow, and Recharge*

INTRODUCTION

Once net precipitation reaches the ground, it either moves into the soil, forms puddles on the soil surface, or flows over the soil surface. Precipitation that enters the soil and is not retained by the soil moves either downward to groundwater or downward and laterally along soil strata to a stream channel or other water body; this is called either *effective precipitation* or *excess precipitation*. Water flowing over the soil surface reaches the stream channel in a shorter time than that flowing through the soil. The allocation of excess precipitation at the soil surface into surface or subsurface flow determines the timing and amount of streamflow that occurs.

The discussion in this chapter shifts emphasis from soil-moisture movement governed by matric potential to water movement over or through the soil that is governed by the force of gravity. We will consider infiltration, percolation, groundwater recharge, and the pathways by which water reaches a stream channel, lake, or wetland.

INFILTRATION

The process by which water enters the soil surface is called *infiltration*, which results from the combined forces of capillarity and gravity acting on water in the soil matrix of micro- and macropores. Micropores relate to the size of pores attributed to soil texture, whereas macropores are larger and occur in soils as a result of several factors discussed later. If water is applied to a dry and unfrozen soil, a rapid initial infiltration rate will normally be observed as a result of the physical attraction of soil particles to water (capillarity) and the flow through macropores that are open to the atmosphere at the soil surface (Fig. 5.1). As a soil reaches field capacity, and if macropores are not prevalent in the soil, the rate of

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

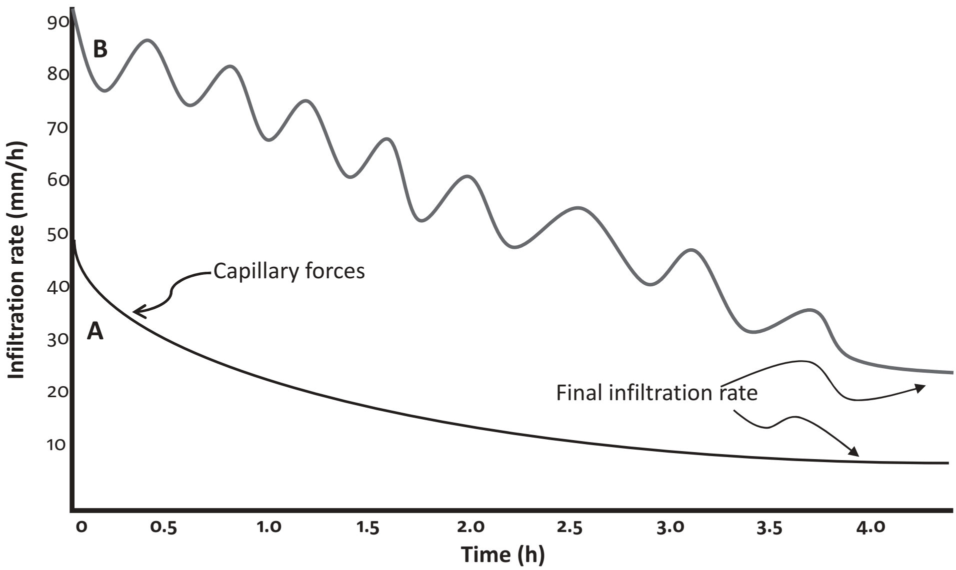


FIGURE 5.1. Infiltration capacity curves for soils **(A)** governed by soil textural porosity and **(B)** governed by macropores that have variable continuity with depth

infiltration typically diminishes exponentially and eventually becomes relatively constant with time (Fig. 5.1). Infiltration at this time is only as rapid as the rate at which the wetting front of water moves through the soil and water drains under the influence of gravity. If macropores are prevalent in the soil, infiltration can be erratic and can vary over time depending on the continuity of macropores (Fig. 5.1). Infiltration rates under the conditions where macropores are prevalent will be discussed later in this chapter. In any case the downward movement of excess precipitation through the soil, called *percolation*, includes drainage from soil horizons in which soil water content exceeds the soil's field capacity or where water flows preferentially through macropores in response to gravity.

The rate at which net precipitation enters the soil surface depends on several soil surface conditions and the physical characteristics of the soil itself. Plant material and litter on or near the soil surface influence infiltration and can be viewed as two hydrologically distinct layers:

1. An upper horizon composed of stems, leaves, and other undecomposed plant material.
2. A lower horizon of decomposed plant material such as litter and duff that behaves much like mineral soil.

The upper layer protects the soil surface from the energy of raindrop impact. When litter is not present, raindrop impact can displace smaller soil particles into pores and effectively seal the soil surface. Plant debris also slows or detains surface runoff, allowing water to infiltrate. The lower duff layer can have a substantial water storage capacity of more than 200% by weight in some instances. Plant litter, therefore, is important as both a water storage component and a protective cover that maintains an open soil surface condition favorable for high rates of infiltration. The conditions that reduce plant litter cover on the soil surface reduce infiltration rates and the infiltration capacity of a soil.

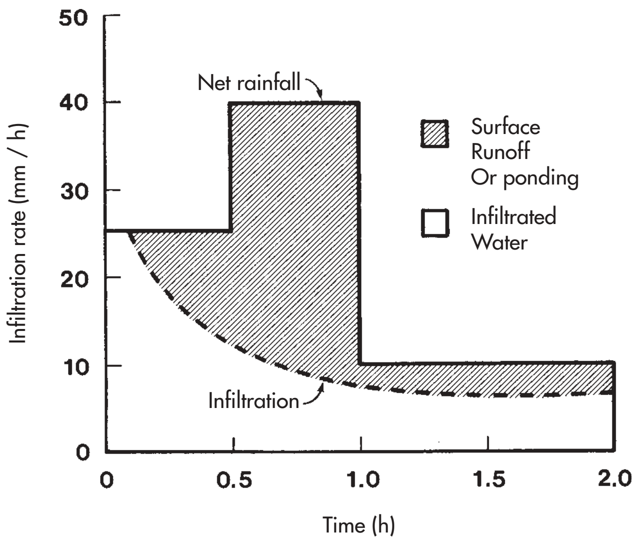


FIGURE 5.2. Relationship between rainfall rate and infiltration rate resulting in surface runoff or ponding

Infiltration Capacity

The maximum rate at which water can enter the soil surface is called *infiltration capacity*. Infiltration capacity diminishes over time in response to several factors that affect the downward movement of the wetting front. The size of individual pores and the total amount of pore space in a soil generally decrease with increasing soil depth. Air entrapment within the pores can temporarily slow infiltration causing erratic infiltration rates (depicted in Fig. 5.1). The swelling of soil colloids can also reduce infiltration rates.

The actual infiltration rate equals the infiltration capacity only when the rate of rainfall or snowmelt equals or exceeds the infiltration capacity. When rainfall or snowmelt rates exceed infiltration capacity, surface runoff or ponding of water on the soil surface occurs (Fig. 5.2). When surface ponding reaches a sufficient depth, the positive pressure of water (head) can result in accelerated rates of infiltration not expected under normal rainfall conditions. Conversely, when rainfall intensity is less than the infiltration capacity, the rate of infiltration equals rainfall intensity. In these instances, water enters the soil and is either held within the soil if soil moisture content is less than the field capacity or percolates downward under the influence of gravity when soil moisture content is greater than the field capacity.

The infiltration capacity of a soil depends on several factors including texture, structure, surface conditions, the nature of soil colloids, organic matter content, soil depth or the presence of impermeable layers, and the presence of macropores within the soil. Macropores function as small channels or pipes within a soil and are nonuniformly distributed pores created by processes such as earthworm activity, decaying plant roots, the burrowing of small animals, and so forth. Soil water content, soil frost, and the temperature of

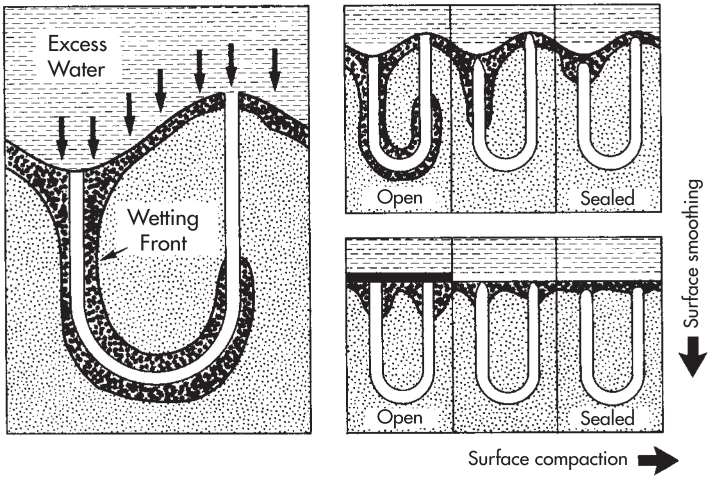


FIGURE 5.3. Effects of surface roughness and surface sealing on infiltration of a soil with a macropore and micropore systems (from Dixon and Peterson, 1971, as reported by Dixon, 1975, ©American Society of Civil Engineers, with permission)

soil and water, all influence the infiltration rate of a soil at any point in time. Land-use activities and vegetation management practices also influence many of the above factors.

The size and interconnections of pores within a soil affect infiltration and the subsequent movement of a wetting front through the soil. As pore size increases, all other factors being equal, the maximum amount of water moving through a pore within the soil is proportional to the fourth power of the radius of the pore:

$$Q = \frac{\pi R^4 \Delta p}{8\eta L} \qquad \textbf{(5.1)}$$

where Q is the volume of water moving through a pore of length L; R is the radius of pore; p is the pressure drop over length L; and η is the viscosity.

This equation, called *Poiseuille's law*, is a fundamental relationship for the laminar flow of water in saturated soils or groundwater systems and explains why water moves at a faster rate through coarse-textured soils than fine-textured soils. Even though matric forces within fine-textured soils exceed those of coarse-textured soils for a given water content, pore size and the interconnected nature of pores ultimately govern infiltration capacity. Similarly, soil structure and the presence of soil fauna and old root channels create a macropore system that increases water conveyance through a soil.

The condition of the soil surface influences infiltration capacities. Surface roughness and the nature of pore openings at the surface largely govern the flow of water into and air out of the soil (Fig. 5.3). Air entrapment reduces infiltration capacity and storage. Soils with rough surfaces have a greater amount of depression storage. Water in depressions is under a positive pressure due to the depth of the water that is greater than atmospheric pressure. Large pores that are open to the atmosphere also promote rapid infiltration–water moves freely into the soil and displaced air freely escapes. Surface compaction or sealing diminishes the effectiveness of large pores that then become barriers to water entry because of the back pressure of air trapped in the sealed pores. Different activities on the soil surface can affect surface and macropore relationships to either enhance or diminish infiltration capacity.

Measurement of Infiltration

The infiltration capacity of a soil can be estimated in the field with *infiltrometers*, the two most common of which are the flooding type and the rainfall-runoff plot type. With either instrument, the entry of water into the soil surface is measured on a small plot of soil. The flooding-type infiltrometer uses a cylinder driven into the soil. Water is added and maintained at a specified depth (usually 10 cm) in the cylinder and the amount of water needed to maintain the constant depth is recorded at specific times. Most often, a double-ring infiltrometer is used in which one cylinder is placed inside another. The inner ring is typically about 30 cm in diameter and the outer ring is 46–50 cm in diameter. Water is added to both cylinders (rings), but measurements are made only in the inner ring. The outer ring provides a buffer that reduces boundary effects caused by the cylinder and the lateral flow at the bottom of the ring. This method is easy to use and relatively inexpensive to apply, but the positive head of water is usually thought to cause higher infiltration rates than might occur during rainfall. Double-ring infiltrometers are useful for obtaining comparisons of infiltration rates for different soils, sites, vegetation types, and treatments.

In the rainfall-runoff plot method, water is applied to the soil surface in a way that simulates rainfall (sprinklers) or natural rainfall events are evaluated. The runoff plot has a boundary strip that forces any surface runoff to flow through a measuring device. Rainfall simulators can be adjusted to represent different drop sizes and rainfall intensities. Rainfall intensities are increased until surface ponding or surface runoff occurs at which time the infiltration capacity has been reached. The rainfall simulator approach is more costly and difficult to apply in remote areas where thick brush or dense trees interfere with sprinkler rainfall simulation. Fewer replications are possible over a given period unless more than one rainfall simulator is available. Infiltration capacities determined by this approach should be more representative of actual infiltration capacities than those determined by flooding-type infiltrometers. However, studies have shown a consistent relationship between infiltration capacities determined by the two methods (Youngs, 1991). Once the relationship is defined, the double-ring approach can be used and the values obtained adjusted with a coefficient to represent more accurate estimates of infiltration.

Infiltration Equations

Infiltration rates have been determined for many soils and plant cover conditions. Typically, the initial phase of infiltration (dry soils) is high and decreases to a relatively constant value as the soil becomes thoroughly wetted (see Fig. 5.1). In the case of agricultural soils and pasture or rangeland conditions, the curves can usually be approximated with several equations of the type described in Box 5.1.

Box 5.1

Infiltration Equations

Horton (1940): $I_t = f_c t + de^{-kt}$

where:

I_t = cumulative infiltration in time t, (cm^3/cm^2)
f_c = constant rate of infiltration after prolonged wetting of the soil (cm/h)
e = base of natural logarithms
d,k = constants

Philip (1957): $I_t = Sp\, t^{1/2} + at$

where:

I_t = cumulative infiltration (cm^3/cm^2) at time t
Sp = "sorptivity" parameter that relates to capillarity or soil matrix forces
a = soil parameter relating to transmission of water through the soil or gravity forces

Holtan (1971): $f_m = ci\, S_a^n + f_c$

where:

f_m = infiltration capacity (cm/h)
c = 0.69 for cm (1.0 for in.)
i = infiltration capacity per unit of available storage (cm/h)
S_a = available storage which is the difference between the potential soil moisture storage and the cumulative infiltration (cm)
n = coefficient that relates to soil texture
f_c = constant rate of infiltration after prolonged wetting of soil (cm/h)

Infiltration measurements of wildland soils, particularly forested soils (as depicted in Fig. 5.1), indicate that infiltration rates change erratically and often do not conform to the smooth curves characterized by the equations in Box 5.1. For nonforested soils, prediction of infiltration can be approximated with models such as DRAINMOD (NCCL, 1986) that use the analytical approach developed by Green and Ampt (Rawls et al., 1983; NCCL, 1986) for infiltration under constant hydraulic head. The *Green–Ampt* equation is essentially a piston flow model that assumes water moves straight down without mixing under a constant hydraulic head following the concept of Darcy's law (Fig. 5.4) and can be expressed as:

$$I_t = k_v(H_o + S_w + L)/L \qquad \textbf{(5.2)}$$

where I_t is the infiltration rate (cm/h); k_v is the hydraulic conductivity (cm/s); H_o is the initial depth of water ponded at or above the surface (cm); S_w is the effective suction at

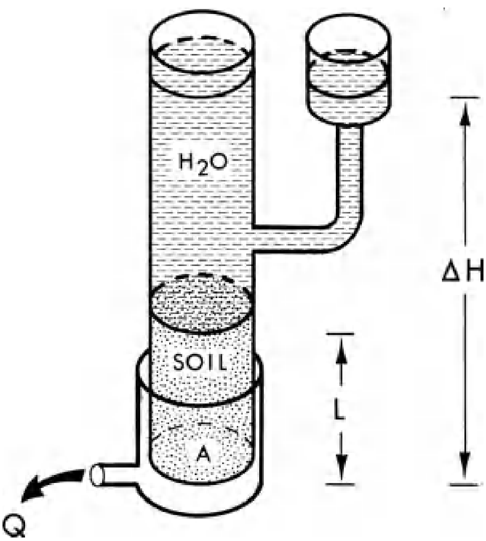


FIGURE 5.4. Darcy's laboratory method of determining the hydraulic conductivity (k_v) of a soil for a given head of water (ΔH) and length of soil column (L) and measuring the quantity of flow per unit time (Q) through area (A)

the wetting front (cm/s); and L is the distance from the ground surface to the wetting front (cm).

Hydraulic conductivity is a physical characteristic of a soil that is constant under saturated conditions. It is best illustrated with Darcy's law:

$$Q = k_v A \frac{\Delta H}{L} \tag{5.3}$$

where Q is the rate of flow (cm^3/s); k_v is the hydraulic conductivity (cm/s); A is the cross-sectional area (cm^2); H is the change in head (cm); and L is the length of soil column (cm).

If the soil-pore system remains unchanged following saturation, Darcy's law approximates the rate of flow. Again, wildland soils with a heterogeneous (nonuniform) pore system might not respond in a manner that is similar to more homogeneous (uniform) media such as large deposits of sand and gravel that serve as aquifers for which the equation is more useful.

There is an accumulation of infiltration that typically occurs with time in soil water models used to account for the soil water portion of a water budget. Various input parameters must be fitted to account for soil moisture content change and the properties of the soil that vary with depth including texture and structure.

Forested soils are usually porous and open at the soil surface, with an extensive macropore system caused by old root cavities, burrowing animals, and earthworms. Initial and final infiltration rates of such soils can be orders of magnitude higher than those of agricultural soils with similar texture. Deep forested soils in humid climates often have infiltration capacities far in excess of any expected rainfall intensity. Surface runoff rarely occurs under such conditions.

A simplified approach for quantifying infiltration has been frequently used in which infiltration (I_t) is estimated by:

$$I_t = A_i + f_c t \tag{5.4}$$

where A_i is the initial loss or storage by the soil (mm); f_c is the final or net infiltration rate (mm/h); and t is the time period after A_i is satisfied to the end of the rainfall event.

Net infiltration rate refers to the relatively constant rate of infiltration that occurs after infiltration has taken place for some time (often 2 hours) and is usually governed by the saturated hydraulic conductivity of a soil unless shallow bedrock or impervious layers are present.

This simplified approach (Equation 5.4) can be used to estimate precipitation excess for stormflow and flood analyses and is particularly well suited for nonurban areas. Precipitation excess (P_e) in millimeters then can be approximated by:

$$P_e = P_n - A_i - f_c t \tag{5.5}$$

where P_n = net precipitation available at the soil surface (mm).

Values of f_c (final or net infiltration rates) have been estimated for a variety of soils, vegetation, and land uses (Table 5.1). Shallow bedrock or impervious layers within a soil can reduce net infiltration rates because of the limited available storage in the soil. Such influences have been reported even for soils under undisturbed forests. For example, an impervious layer at 0.2 m depth caused the upper soil layer to become quickly saturated during rainstorms in a tropical forest in Australia (Bonell et al., 1982). This resulted in surface runoff even though the pores of the undisturbed soils were apparently open at the surface because the soil pores became filled with water above the impervious layer that was then displaced back to the surface.

Land-use Impacts on Infiltration

The activities that compact or alter the soil surface, soil porosity, or the vegetative cover can reduce the infiltration capacity of a soil. Driving vehicles or pulling logs over a soil surface, intensive livestock grazing, and intensive recreational use can compact the surface and reduce infiltration. Exposing a soil to direct raindrop impact will also diminish the openness of the surface soil and reduce infiltration capacities.

Harvesting timber in a 110-year old Douglas-fir stand in Oregon with a low ground-pressure, torsion-suspension skidder resulted in 25–45% increases in soil bulk density to a

TABLE 5.1. Net (or final) infiltration rates for unfrozen soils

Soil category	Bare soil	Row crops	Poor pasture	Small grains	Good pasture	Forested
			(mm/h)			
I	8	13	15	18	25	76
II	2	5	8	10	13	15
III	1	2	2	4	5	6
IV	1	1	1	1	1	1

Source: From Gray (1973), with permission.

I, coarse-medium texture soils over sand or gravel outwash; II, medium texture soils over medium texture till; III, medium and fine texture soils over fine texture till; IV, soil over shallow bedrock.

depth of 15 cm (Sidle and Drlica, 1981). The area affected by skidding amounted to 13.6% of the total area logged plus 1.5% of the area used as a landing. The greatest compaction resulted from frequent travel over wet soils.

The compaction of surface soils by yarding and skidding of logs reduces infiltration capacities and can result in surface runoff and erosion. If soils are allowed to recover, the effects of such operations are usually negligible after 3–6 years except where soils were heavily disturbed (Johnson and Beschta, 1980).

Livestock grazing can reduce infiltration capacity by removing plant material, exposing mineral soil to raindrop impact, and compaction of the surface. Surfaces compacted by intensive grazing can reduce infiltration capacities over a wider area than can activities such as skidding logs. Infiltration capacities for different grazing and vegetative conditions in Morocco are compared in Table 5.2. Differences in soils and vegetative cover often confound comparisons of infiltration capacities among different grazing conditions. As a rule, rangelands in good to excellent condition with light grazing exhibit infiltration capacities at least twice those of rangelands in poor condition with heavy livestock grazing.

Land use can also affect infiltration capacities indirectly by altering soil moisture content and other soil characteristics.

Water-Repellent Soils

Dry soils normally have an affinity for adsorbing liquid and vapor water because of the strong attraction between the mineral soil particles and water (capillarity). Water droplets that are placed on a wettable soil surface are absorbed immediately. However, when water droplets are placed on the surface of a dry water-repellent (hydrophobic) soil, they tend to "bead up" and not penetrate the soil because the mineral particles are coated with hydrophobic substances that repel water (see Fig. 2.1). Fires that occur in several types of vegetative communities can drive these hydrophobic substances into the soil, forming a hydrophobic layer below the soil surface. The presence of water-repellent soils reduces infiltration rates in a similar manner to that of a hardpan layer that restricts water movement through the soil (Fig. 5.5). Infiltration of water into a water-repellent soil is inhibited and often completely impeded, in which case the net precipitation reaching the ground quickly becomes surface runoff or shallow lateral flow. How the infiltration rate changes

Table 5.2. Comparisons of soils and infiltration relationships (double-ring infiltrometers) for three land-use conditions in northern Morocco

	Heavily grazed, Doum palm vegetation	Moderately grazed, brushland	Ungrazed, afforested (Aleppo pine)
Soil texture	Coarse	Medium	Fine
Soil organic matter content (%)	1.47	1.77	2.7
Soil bulk density (g/cm^3)	1.44	1.42	1.22
Vegetative cover (%)	12.5	41.3	99
Slope (%)	0–10	5–25	5–40
Initial infiltration rate (mm/h)	179	194	439
Infiltration rate after 2 hours (mm/h)	43	65	226

Source: Adapted from Berglund et al. (1981).

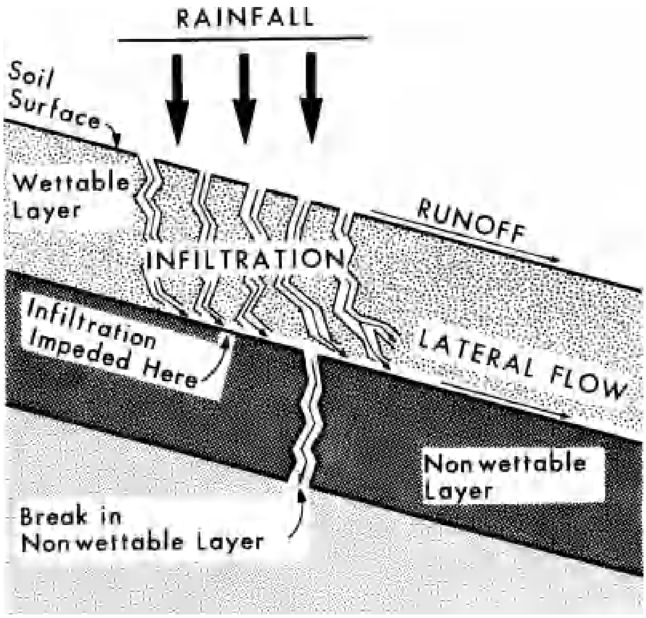


FIGURE 5.5. Effect of water-repellent layer on infiltration and surface runoff (from DeBano, 1981).

because of a water-repellent layer is illustrated in Figure 5.6. Over time, the water-repellent compounds can break down and the soil can once again achieve nonhydrophobic infiltration characteristics.

Water repellency appears to be caused by several mechanisms involving the presence of organic matter consisting of long-chain hydrocarbon substances coating the mineral soil particles. These mechanisms include:

- irreversible drying of the organic matter such as the difficulty encountered when rewetting dried peat (organic) soils;
- plant leachates that coat mineral soil particles, for example, coarse-grained materials easily made water repellent by plant leachates;
- coating of soil particles with hydrophobic microbial byproducts, for example, fungal mycelium;
- intermixing of dry mineral soil particles and dry organic matter; and
- vaporization of organic matter and condensation of hydrophobic substances on mineral soil particles during fire, that is, heat-induced water repellency.

Water-repellent soils are found throughout the world on both wildland and agricultural croplands and have been a concern to watershed managers since the early 1900s. The nature of the water repellency of soils on the sites supporting chaparral shrubs in southern California has received widespread attention since the 1960s (DeBano, 2000a). Organic matter that accumulates in the litter layer under chaparral shrubs is leached to the soil. Water repellency then results as the organic substances accumulate and mix in the upper soil profile. The naturally occurring water repellency in soils is intensified by the frequent

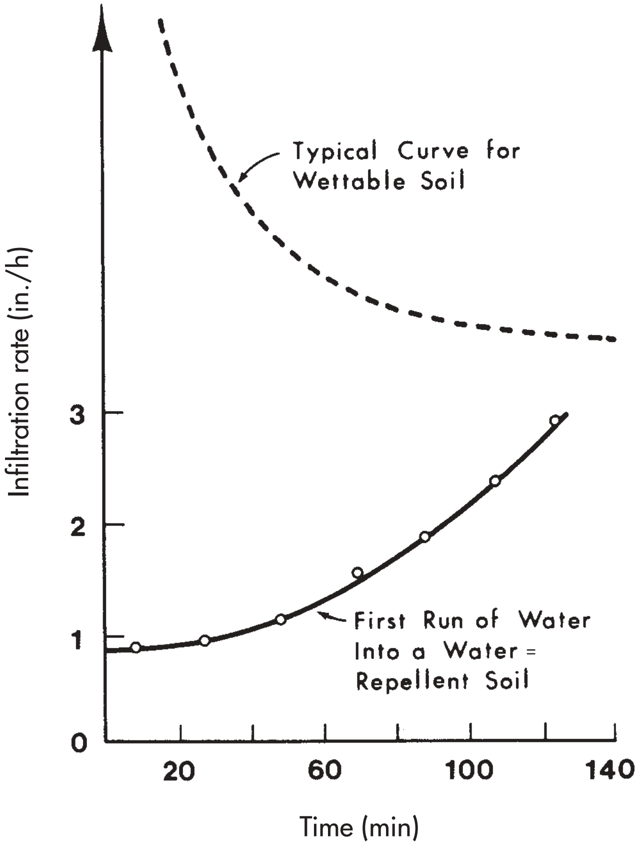


FIGURE 5.6. Infiltration rates of water-repellent soil (shown in Figure 5.5) and wettable soil (from DeBano, 1981)

wildfires in the region, which volatilize the organic substances, driving the water-repellent layer deeper into the soil (Box 5.2).

Wetting agents or *surfactants* have been used in treating fire-induced water repellency to minimize their effects of infiltration. However, whereas non-ion surfactants have been somewhat effective in reducing runoff and erosion from small plots, they are often ineffective on a watershed basis. Because the magnitude of fire-induced water repellency is related to fire severity, a remedial approach on a watershed is to prescribe controlled low-severity fire to reduce fuel accumulations that lead to uncontrolled wildfire conditions and the resulting severe water repellency.

There is also a widespread concern about the effects of natural water repellency on the productivity of agricultural croplands. About 5 million hectares of agricultural and pasture landscapes in Australia and New Zealand have been adversely affected by water repellency, resulting in large financial losses for the landowners (Blackwell et al., 1993). Remedial treatments to mitigate this problem include applications of wetting agents, direct drilling, wide-furrow sowing, and the use of microorganisms and fertilizers to stimulate microbial

Box 5.2

The Role of Fire and Soil Heating in Intensifying Water Repellency (from DeBano, 2000b)

Field and laboratory studies have confirmed that water repellency is intensified by soil heating during a fire. The combustion and heat transfer during a fire produces steep temperature gradients in the surface layers of the mineral soil. Temperatures in the canopy of burning chaparral shrubs in southern California can be more than 1100°C, whereas temperatures reach about 850°C at the soil–litter interface. However, temperatures at 5 cm deep in the mineral soil probably do not exceed 150°C because dry soil is a good insulator. Heat produced by combustion of the litter layer on the soil surface vaporizes organic substances, which are then moved downward in the soil along the steep temperature gradients until they reach the cooler underlying soil layers where they condense. Incipient water repellency at different soil depths is also intensified in place by heating, because organic particles are heated to the extent that they coat and are chemically bonded to mineral soil particles. Movement of hydrophobic substances downward in the soil occurs mainly during the fire.

After a fire has passed a site, the continued heat movement downward through the soil can revolatilize some of the hydrophobic substances, resulting in thickening the water-repellent soil layer or fixing the hydrophobic substances *in situ*. The final result is a water-repellent layer below and parallel to the soil surface on the burned area which restricts water movement into and through the soils.

breakdown of water repellency. Mixing large amounts of clay in the upper water-repellent layer has also been used.

Undisturbed peat (organic) soils are porous and exhibit high infiltration capacities, but a hydrophobic condition has been observed when they have been mined or allowed to become dry (Schroeder, 1990). Mining of peat for horticulture exposes large areas of peat soil. Once the upper layer of sphagnum is removed and the darker, more decomposed peat becomes exposed to direct solar radiation and wind, the soil surface can dry to a point where it becomes water repellent. Surface runoff can increase after this occurs. While the cause of this hydrophobic condition is unknown, it could be similar to that induced by fire. The easiest way to alleviate this condition is to manage water levels to prevent the peat surface from becoming excessively dry.

Soil Frost

Soil frost is common during winter and spring in cold continental climates. It can also occur periodically in milder climates and can lead to serious flooding, particularly when

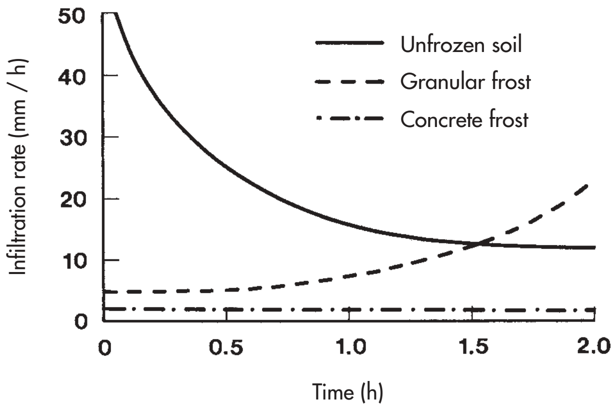


FIGURE 5.7. Effects of soil frost on infiltration rates (adapted from Gray, 1973)

snowmelt and high-intensity rainfall occur on frozen soils. The influence of soil frost on infiltration capacity is determined largely by the moisture content of the soil when it freezes. A saturated soil upon freezing can act as a pavement with little infiltration, a condition referred to as concrete frost (Fig. 5.7). If the soil is not saturated when freezing occurs, a granular, more porous frost develops resulting in a better infiltration capacity than found with concrete frost. Under conditions of granular frost, some melting of soil frost occurs as infiltration continues. The soil pores in effect then become larger and able to transmit water at a faster rate. This explains the increase in infiltration over time.

The occurrence of soil frost is affected by vegetative cover, soil texture, depth of litter, and depth of snow. Snow and litter act as insulators, that is, the deeper the snow before freezing, the less likelihood of soil frost. By altering vegetative cover, particularly forest cover, frost can be affected indirectly because of the influence of the forest canopy on snowpack accumulation and distribution. In general, removal of forest cover results in more frequent and deeper occurrences of soil frost. Compaction of soil surface horizons also can increase the depth of frost penetration. The effects of different forest cover conditions on soil frost in northern Minnesota are illustrated in Figure 5.8. The balsam fir stand intercepts more snow than hardwood stands, resulting in less snow depth and deeper frost than under the hardwood stands. Unforested open areas, and particularly cultivated soils that are exposed during cold winters, experience the deepest frost penetration that is often concrete frost.

PATHWAYS OF WATER FLOW

Infiltrated water that is not retained in the soil can take several pathways as depicted in the hydrologic cycle (Chapter 2). This water can drain downward through unsaturated soil and escape capture by plant roots, ultimately reaching groundwater or saturated soil layers above the water table that intercept and divert water laterally to stream channels, lakes, and wetlands. Saturated soil layers occur when the rate of percolation downward becomes impeded by a confining layer with limited permeability. Water that reaches shallow groundwater can also be discharged into surface water bodies where the water table

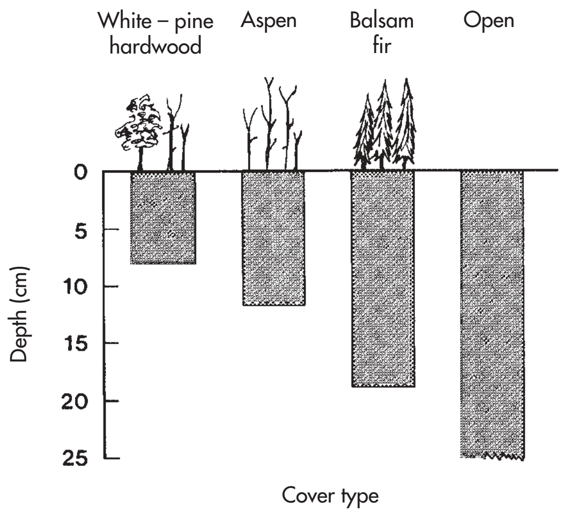


FIGURE 5.8. Depth of soil frost during midwinter (January to mid-March) on loamy soils in northern Minnesota under different cover types; the vertical scale is exaggerated (from Weitzman & Bay, 1963)

comes in contact with those water bodies at or near the soil surface. In contrast, vertical seepage of water can reach deep groundwater aquifers whose boundaries do not coincide with surface watershed boundaries. Such seepage represents a loss of water from the watershed. This section discusses these various pathways of water flow and how they affect groundwater recharge and streamflow. Groundwater–surface water interactions with lakes, riparian zones, and wetlands are discussed in Chapter 7.

Groundwater Recharge

Groundwater recharge is a hydrologic process where water typically percolates (drains) downward from the earth's surface to groundwater. This process usually occurs in the *vadose zone*, defined as the zone below plant roots and above the water table (Fig. 5.9). However, in some shallow groundwater systems groundwater recharge can occur above the vadose zone. Recharge occurs both naturally (through the water cycle) and artificially where rainwater and or reclaimed water is routed into the subsurface.

Groundwater recharge can also occur when water moves from a lake or wetland edges into groundwater. This process is defined as *outseepage*. Outseepage occurs where a continuous saturated hydraulic head from a lake or wetland causes water to flow from the surface water into a groundwater aquifer. This flow can only occur when there is no impeding layer between the surface water body and groundwater (Fig. 5.9).

A third type of groundwater recharge occurs from the streams that are perched above a groundwater aquifer. Because the flow in such streams is not sustained by groundwater discharge, they do not flow throughout the entire year. Streamflow can be sustained for periods of time by either rainfall or snowmelt, but during periods with zero or limited precipitation streamflow can cease as water seeps from the channel into the ground. This condition is defined as a *losing reach* and is common in drylands (Fig. 5.9).

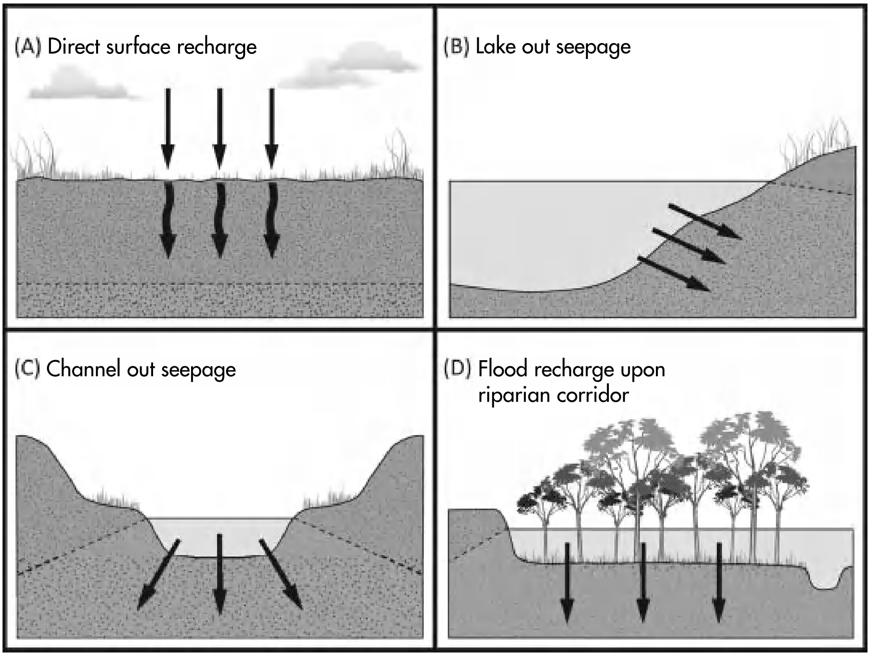


FIGURE 5.9. Types of groundwater recharge found in varying climatic and geologic settings.

A fourth type of recharge that can occur within riverine systems is *bank storage*. Bank storage occurs when the flow of water in the channel rises to an elevation higher than the water table of the adjacent aquifer that is discharging into the river channel. Typically, flood waters from the upper watershed are delivered to the lower watershed valley. Water percolates from the river into the bank and into the lower elevation water table of the valley sediment adjacent to the river. Recharge from bank storage is greatest during periods of high flows near or above the top of the riverbank, called bankfull. Typically, the process occurs during high flood stage where upgradient water is delivered to a downgradient floodplain (see Chapter 16 example from South Africa). Streamflow that exceeds bankfull moves into a downstream floodplain and riparian corridor where it percolates into an underlying alluvial aquifer and does not return to the river as surface flow (Fig. 5.9).

Groundwater Recharge Zones and Land Use

Upland forested watersheds are commonly viewed as being important recharge zones for aquifers because forests occur in the areas with high annual precipitation and are associated with the soils that have high infiltration capacities. If true, then is groundwater recharge affected by land management activities such as forest harvesting, regeneration of forest cover, and conversion of forests to annual crops or other types of vegetative cover? In considering this question, we will first examine the processes affected and then the implications of such changes for groundwater supplies. We will consider how different actions on upland recharge areas can affect the processes of groundwater recharge and discharge.

Any actions in recharge areas that increase the amount of water in soil storage can increase the amount of water that is available for groundwater recharge. Actions such as road construction or urbanization of uplands can reduce infiltration capacity that leads to increased surface runoff. If such disturbance is widespread, surface runoff can be increased at the expense of subsurface and groundwater flow. The conversion of forests to urban development, agricultural croplands or pastures could result in a more widespread and permanent impact on infiltration capacity and recharge pathways. The net effect of such activities depends on whether reductions in evapotranspiration (*ET*) or reductions in infiltration have the greatest impact on recharge.

Although not well documented through controlled watershed experiments, it is possible that widespread soil disturbance in a recharge zone could cause groundwater-fed perennial streams to become dry during seasonal low-flow periods. An example of this process occurs in south central Minnesota because the shallow groundwater is removed by extensive subsurface drainage. Such occurrences would be rare in most landscapes; however, and significant only where small catchments feed a localized groundwater aquifer. Otherwise, where recharge areas are vast, the opportunities for water to recharge a groundwater aquifer are too great.

Most small-scaled controlled watershed experiments have shown increases in recession flow and baseflow when changes on the watershed reduce *ET* losses, suggesting an increase in groundwater recharge. Such has been the case following timber harvesting or conversion from forest cover to vegetative cover with lower annual consumptive use of water described in Chapter 12.

Methods of Estimating Groundwater Recharge

Rising water demands and increased water scarcity for both humans and nature make the need for improved regional recharge estimates critical for transitioning to sustainable water resources management (Barlow, 2002). Recharge is generally the most difficult component of the groundwater system to quantify (Bredehoeft, 2007) largely because of the difficulty in estimating *ET*, infiltration, and other processes across the landscape that determine the water budget. Soil surveys that have been developed across the USA provide information about the soil properties and the spatial location of the soil types that can be used to roughly approximate spatial differences in infiltration and soil permeability. No such surveys exist to approximate *ET*. Because of the difficulty of measuring recharge, hydrologic models are often used to estimate recharge.

Many hydrologic models ignore spatial and temporal variations in recharge rates because of limited available data or the method is not adequate to accurately evaluate the variations at the scales of interest (Hyndman et al., 2007). There are, however, a number of methods for estimating the spatial distribution of recharge across a landscape. These methods can be categorized into (1) physically based and (2) regression methods.

The most common way to estimate groundwater recharge over large terrestrial landscapes, excluding lake and wetland zones of recharge, is by the application of physical models using the principles of soil physics to estimate recharge. Physical models are those that attempt to actually measure the volume of water passing below the root zone. Indirect physical methods rely on the measurement or estimation of soil physical parameters, which, along with soil physical principles, can be used to estimate the potential or actual recharge.

Other indirect measures would use geochemical or isotopic tracers at selected recharge locations to validate the soil physics expressed through a numerical model. Typically, these models use varying equations previously discussed in this chapter to account for water movement through the vadose zone that provides a value for deep percolation or recharge.

After months without rain, the level of the rivers in humid climates is low and the discharge would presumably be derived exclusively from drained groundwater. Under such conditions, recharge can be calculated from baseflow discharge if the catchment area is known. An application of this approach is presented in Box 5.3.

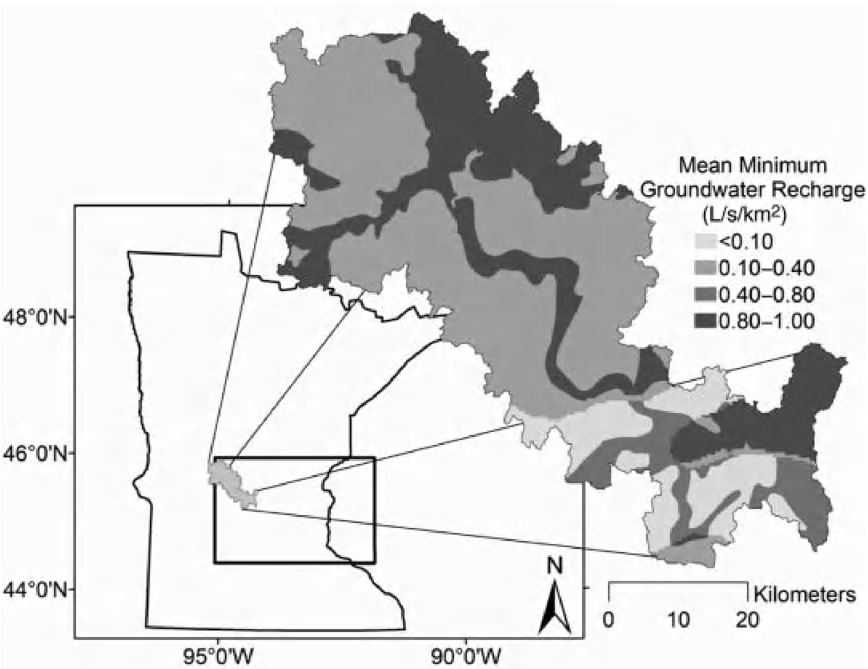
Box 5.3

Exploring a Watershed-based Approach to Groundwater Recharge (Used with Permission from Peterson, 2011)

A physically based model, the Watershed Characteristics Approach (WCA), was developed to link the spatial variation of the complex interactions between multiple watershed landscapes and groundwater recharge. Given that watersheds are somewhat self-organizing systems depending on the intensity of management, their characteristics are a result of adaptive, ecological, geomorphic, or land-forming processes. Therefore, a watershed in a given climate and geologic setting may establish the patterns that upon examination can lead to a simplification of descriptions used in analysis and predictions (Sivapalan, 2005). By assuming that the groundwater divide coincides with the surface topographic divide used to delineate the watershed drainage area, the WCA is a system-based recharge estimation model that uses regionalization (Pinneker, 1983) of watershed landscape characteristics to define unique hierarchical hydrogeological units and establish their link to hydrologic characteristics (e.g., stream discharge, groundwater recharge) (Kanivetsky and Shmagin, 2005).

Hydrologic regionalization is used to determine hydrologically similar units, including soils, topography, geology, climate, hydrology, and vegetation. This regionalization works at a multiscale level to account for the characteristic hydrologic response of an entire unit. The hydrologic response of this unit is based on measurements at an appropriate scale, which can be directly related to the hydrologic response of geometrically similar hydrologic units elsewhere.

Low-flow data from United States Geological Survey (USGS) gauges in the Sauk River near St. Cloud, Minnesota, were used to estimate minimum groundwater recharge rates for different locations. WCA predicted recharge rates from 1955 to 1978 for the various subwatersheds in which low-flow data were available (see Figure below). The minimum recharge rates determined by the WCA were lower that the mean annual recharge rates calculated by Lorenz and Delin (2007) which would be representative of periods with higher rates of recharge. Nevertheless, the WCA approach demonstrates that watershed landscape units can be spatially linked to predict groundwater recharge during low-flow periods. Such watershed-based approaches should be enhanced through improved technology and expanded monitoring.



Box 5.3. Minimum groundwater recharge rates predicted by the WCA for the Sauk River, Minnesota, from 1955–1978 (Peterson, 2011, ©University of Minnesota, with permission)

Physically based models, to various degrees of complexity, apply equations that quantitatively describe the various processes involved in the water balance for the land surface and shallow unsaturated zone. Because it is difficult to estimate spatially dependent recharge with uncertain or incomplete data, physically based models rely on calibration to account for the lack of knowledge arising from the uncertainty of actual processes and interactions. Compounding uncertainty makes this approach an estimated guess using the best professional judgement of the modeller. In contrast, when a regression method is used, it is hoped that the response of the hydrologic system can be predicted by separating out the individual components of a system; however, the response of a hydrologic unit is not simply the linear combination (regression) of its component parts. When used to model regional recharge across a large scale, this traditional estimation approach can fail to adequately acknowledge the relationships which exist between the landscape (surface and interior) and hydrologic response (Nathan and McMahon, 1992; Smakhtin, 2001; Kroll et al., 2004; Stepinski et al., 2011). As explained in Box 5.3, the approach by Peterson (2011) links the spatial variation of the landscape with the hydrologic response.

Water Flow into Stream Channels

Water can enter stream channels from a myriad of flow pathways. Several factors determine which pathways are followed, including landscape position, soil type, vegetative cover, land use, landscape position and geologic conditions underlying the watershed, and the intensity and rates of rainfall or snowmelt over the watershed. Depending on which pathways are taken, the journey of water through some routes to the channel can be very brief, whereas other routes can take long time periods – sometimes years before reaching a stream channel. Water flows into a channel from the upper to lower reaches with the cumulative volume of water in the channel increasing. At the outflow of any watershed, the magnitude of flow through each of these pathways along the channel continuum determines the integrated streamflow response over time. We discuss the pathways of flow and the corresponding response of streamflow in the following section. The methods of measurement, analysis, and modeling of streamflow are discussed in Chapter 6.

Streamflow Regimes

The pathways by which water enters a stream determine the seasonal and annual flow patterns of a stream, generally referred to as streamflow regimes. *Streamflow regimes* can be perennial, intermittent, or ephemeral in their persistence. A *perennial* stream flows throughout a year with the possible exception of severe drought conditions; an *intermittent* stream flow in the wet season but ceases to flow in other seasons of a normal year; and an *ephemeral* stream flows in direct response to a rainfall or snowmelt event. Ephemeral streamflow is most common in dryland environments, although it can also occur in other climates. Various processes and varying pathways determine how much and how rapidly the precipitation falling on a watershed will augment the flow of a perennial stream, add to the flow of an intermittent stream, or initiate the flow of an ephemeral stream.

It is helpful to think in terms of *storage* and *conveyance* when trying to understand how watershed and stream channel conditions affect a streamflow regime. A large amount of precipitation can be stored in the soil or surface water bodies and then released into the atmosphere by *ET*. The integrated response of streamflow following rainfall or snowmelt for a particular watershed or river basin is best described with a hydrograph.

Streamflow Hydrograph

The graphical representation of streamflow discharge over time is a *hydrograph* (Fig. 5.10). Hydrographs are widely used by hydrologists to study the integrated streamflow response of a watershed or river basin to various climatic conditions, meteorological events, and land use changes. The amount of precipitation that flows via the different pathways of a watershed determines the interstorm flow and the shape and peak flow magnitude of the stormflow hydrograph.

Baseflow

Typically groundwater feeds a *perennial stream*, one that flows continuously throughout a year, represented as pathway D in Figure 5.10. In addition to groundwater, long-term subsurface drainage from uplands can sustain streamflow over long periods between rainfall or snowmelt events – this component of streamflow is called *baseflow.* This graphically

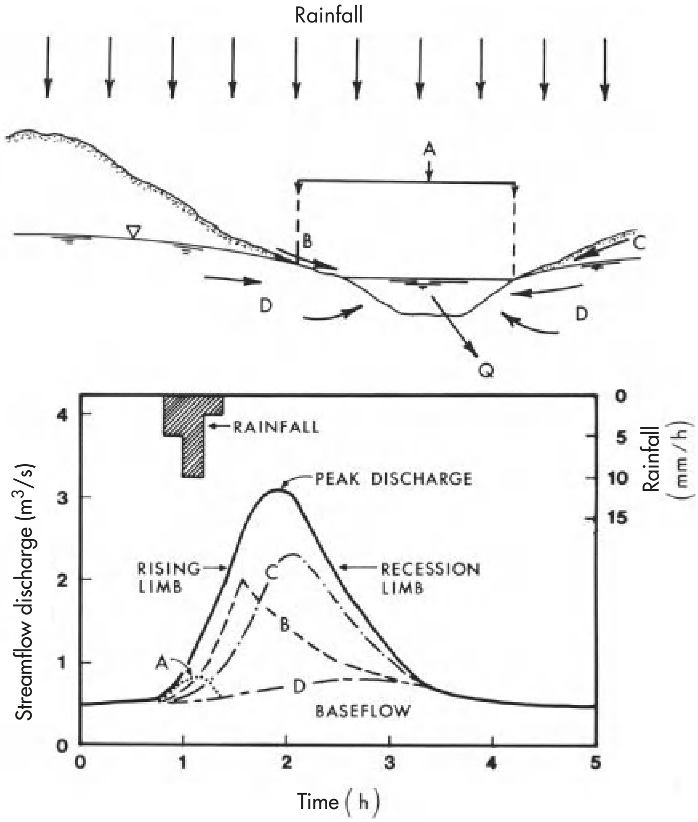


FIGURE 5.10. Relationship between pathways of water flow from a watersheds and the resultant streamflow hydrograph. A, channel interception; B, surface runoff; C, subsurface flow or interflow; D, groundwater or base flow; *Q*, streamflow discharge

level part of hydrograph does not typically respond quickly to excess precipitation because of the long and tortuous pathways through which water flows.

Stormflow. Following rainfall storms or periods of snowmelt, the hydrograph rises in response to the quicker responding pathways of flow to the stream channel. The most direct pathway is precipitation that falls directly on the stream channel and adjacent saturated areas, called *channel interception* (A in Fig. 5.10), causing the initial rise in the streamflow hydrograph; it ceases after precipitation stops and flows to the outlet of the watershed.

Surface runoff, also called *overland flow*, is the water that flows over the soil surface as the result of flows from impervious areas on the landscape (rock outcrops, road surfaces, etc.), locally saturated areas, or from the areas where the rainfall rate exceeds the infiltration capacity of the soil (B in Fig. 5.10). Some surface runoff is detained by the roughness of the hillslope surface and, therefore, can be slowed where it can infiltrate at some downslope location before reaching a stream channel. This pathway results in flow reaching the channel later than channel interception but quicker than groundwater. *Interflow* is subsurface flow that is part of excess precipitation that infiltrates but arrives at the stream channel over a short enough period to be considered part of the storm hydrograph (C in Fig. 5.10). This flow can be rapid in soils with large, connected "macropores" and is considered the major pathway in most forested watersheds with deep soils.

The sum of channel interception, surface, and interflow is called *stormflow*, *direct runoff*, or *quickflow*. The term stormflow is used in this book to describe this part of the hydrograph. Stormflow is graphically depicted as having a rising limb, a peak, and recession flow – all of which are above baseflow in a perennial stream. The stormflow volume is the sum of flow that is above baseflow and is directly attributed to a particular rainfall or snowmelt event.

Whereas the four major pathways of water flow might be visualized conceptually, separating one from the others and measuring each pathway is difficult. Rainfall or snowmelt can follow a combination of surface and subsurface routes before reaching the stream channel. Water can infiltrate in one area of a watershed and exfiltrate, that is, return to the surface, downslope and flow over the land surface for a distance. Conversely, some surface runoff can collect in depressions on a hillslope surface to be evaporated or infiltrated later. The stormflow hydrograph depicts the excess water from rainfall or snowmelt amounts that does not become stored in the watershed.

Saturated overland flow can occur when the water already contained in the soil or riparian system (prestorm event) rises to the surface and runs into the channel. The differences between "new precipitation" hitting the surface and running off as surface runoff and saturated overland runoff have been measured by geochemical and isotopic techniques that are discussed in Chapter 16. Surface runoff is a relatively large component of the stormflow hydrograph for impervious urban areas, but it is insignificant for forested areas with well-drained deep soils and elastic aquifers.

If water does not infiltrate where it contacts the surface, quickflow in one part of a watershed can infiltrate at some downgradient location before reaching a stream channel. This pathway results in flow reaching the channel later than overland runoff but theoretically quicker than subsurface pathways.

Interflow is considered the primary flow pathway in most well-drained forested watersheds. However, in some instances based on geochemical and isotopic studies (Kendall and McDonnell, 1998), interflow could be more aptly called *shallow groundwater* flow as opposed to *deep groundwater* flow. Once new water (from recent precipitation) enters the groundwater and mixes with pre-existing water, it assumes a new geochemical signature that represents a composite mixture reflective of the new source water. Studies have shown that different hydrologic pathways produce different isotope or geochemical signatures (Magner and Alexander, 2008; Stets et al., 2010). For example, water outseeping from a lake into a deep groundwater aquifer will show unique lake-water attributes compared to snowmelt recharging into shallow groundwater that will contain unique snow attributes (Stets et al., 2010). Lake water will develop a unique isotopic signature if allowed to evaporate, whereas snow typically shows no evaporative signature. Further, snow is composed of lighter isotopes of hydrogen and oxygen. Evaporated lake water contains more heavy oxygen and heavy hydrogen because the heavy hydrogen evaporates faster than the heavy oxygen (Kendall and McDonnell, 1998). Differences in isotopic signatures are discussed further in Chapter 16.

Deep groundwater is typically considered the main source of water to streamflow during minimum flow conditions. Low flow occurs in periods of droughts when the only sustaining flow usually comes from older deep groundwater contained in a vast regional aquifer. There is evidence to suggest that in some geologic settings, deep groundwater can contribute to stormflow where shallow groundwater is “strongly” linked to deeper portions of an aquifer (Kendall and McDonnell, 1998). This process is defined as *hydraulic displacement* or *translatory flow* where all the subsurface water is hydraulically connected and moves like dominoes from recharge to discharge (D in Fig. 5.10 could be higher than shown). This observation implies that old, prestorm-event groundwater can be pushed into a stream channel nearly as rapid as interflow. Consider a garden hose used during a sunny summer morning but not used again until late afternoon for a cold drink. If you were hot and thirsty, would you drink the first flush of water from the hose? Probably not, because water stored in the hose during the day was heated by the sun. You would more likely wait for the warm water stored in the hose to be translated or displaced from the hose until cold water from the house reached the end of the hose. This same hydraulic connection exists in nature.

Stormflow and flood studies do not normally attempt to separate the pathways of flow; however, some insight into possible major pathways of flow can be inferred by examining the stormflow response to large rainfall or snowmelt events in the context of the magnitude, intensity and duration of rainfall, and the watershed antecedent moisture conditions. Based on the preceding discussion, the separation of a hydrograph in terms of a time response can be related to the major flow pathways that constitute streamflow response to storms. Hydrologic models – to varying degrees – attempt to represent the major flow pathways to predict the integrated streamflow response in the form of a hydrograph.

The array of pathway options makes it difficult to manage nonpoint source pollution that occurs from across a watershed. This is why the concept of *hydraulic residence time* is important in determining pollutant fate and transport management. Understanding transport, storage, and lag times associated with water and solute movement aids decision makers in selecting the appropriate management actions required to restore or protect water resources as discussed in Chapter 14.

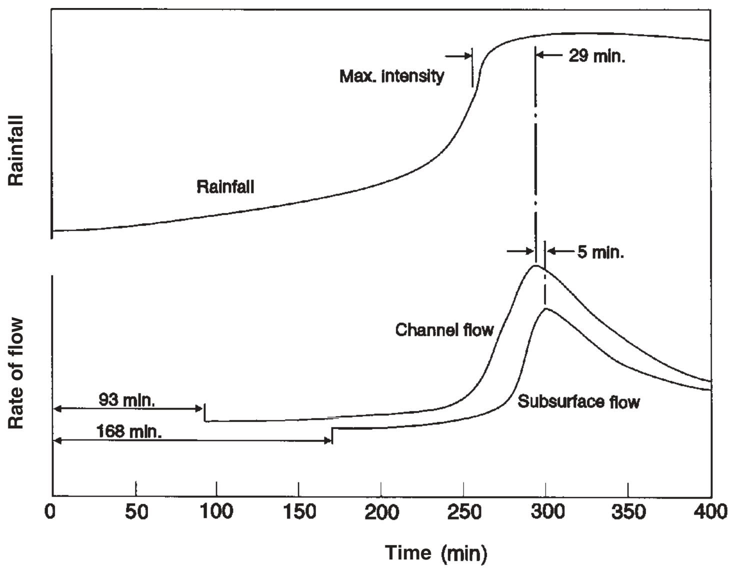
Hillslope Hydrology

At times, the hydrology of hillslopes are studied rather than entire watersheds to better help us understand the processes of surface runoff and interflow in generating streamflow at a more easily measurable scale. Hillslopes also represent the smallest unit on which most of the processes occur that one would expect to see in a watershed (Troendle, 1985). The fate of precipitation with respect to interception, infiltration, soil moisture storage, surface runoff, and percolation can be observed within a hillslope. The pathways through which water flows are largely determined by infiltration capacities and the characteristics of soil strata on hillslopes. In many areas with deep soils and supporting forest vegetation, the runoff at the toe of a slope can largely be the product of subsurface flow through the soil profile rather than quickflow. As infiltrated water moves through the soil profile, constricting layers are encountered and saturated zones develop. The saturated zone develops a head of pressure along hillslopes that can force water to move along soil layers or through the soil profile, eventually reaching the toe of the slope. When the zone of saturation is continuous in the profile, any addition of water at the top of the zone results in an increased rate of flow at the toe of the slope as described above and referred to as translatory flow. When large continuous pores (macropores) are present, water can move through the soil rapidly, as through a pipe (recall Equation 5.1). This movement is called *pipeflow*.

Box 5.4

Stormflow Response of Hillslopes and Small Watersheds (from Beasley, 1976)

Runoff from plots (less than 0.1 ha) on forested hillslopes in Mississippi produced negligible overland flow and negligible shallow subsurface flow (above the B-horizon) from rainfall events. Streamflow peaks from 1.86 and 1.62 ha watersheds, averaged more than 36 storms, responded within 5 minutes of peak subsurface flow from the plots (see hydrograph below). It was concluded that flow through macropores (pipeflow), and not translatory flow, could be the only explanation for such a quick response.



Relationship between timing of rainfall and corresponding subsurface flow and channel flow responses averaged for 36 storms for hillslopes in Mississippi (Beasley, 1976, as presented by Troendle, 1985).

Hillslope studies, particularly in forested regions, indicate that stormflow production can be largely the product of subsurface flow (Box 5.4).

Variable Source Area Concept

Wildland watersheds are heterogeneous combinations of vegetative covers, soils, and land uses with varying topographic features and stream channel configurations exhibiting a wide

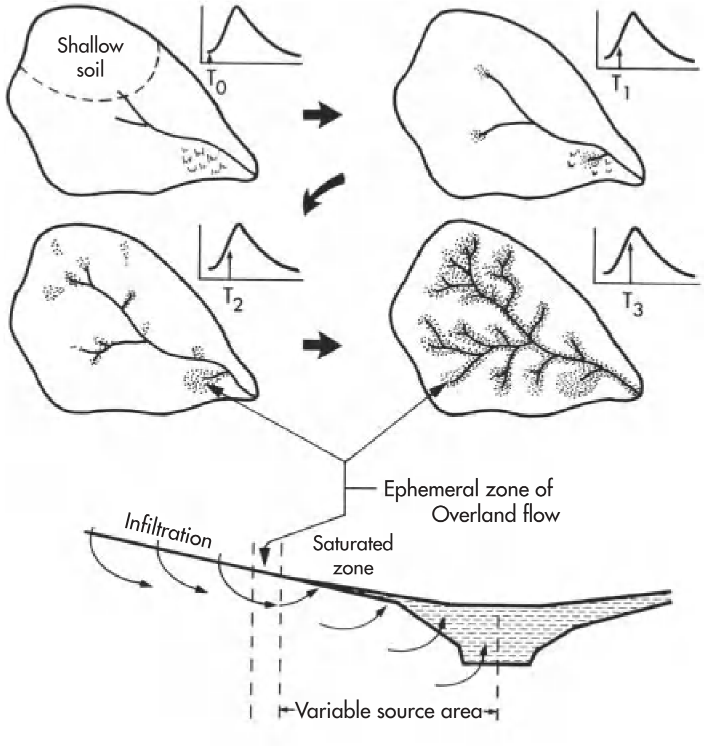


FIGURE 5.11. Schematic of the variable source area of stormflow and the relationship between overland flow of water and the zone of no infiltration. The small arrows on the hydrographs indicate streamflow-response changes as the variable source area expands (modified from Hewlett and Troendle 1975; Hewlett, 1982, ©University of Georgia Press, by permission)

responses to rainfall or snowmelt events. At one extreme, forested watersheds with deep, permeable soils can have high infiltration capacities and exhibit predominantly interflow below the soil surface. On the other extreme, rangelands with shallow soils can have low infiltration capacities and exhibit a quick or flashy streamflow response dominated by surface runoff. Perhaps the most common situation is a watershed where some areas such as rock outcrops and roads produce surface runoff for almost any rainfall event, and other areas seldom, if ever, produce surface runoff.

The *variable source area concept* (VSAC) explains the mechanisms of stormflow generation from watersheds that exhibit little surface runoff (Hewlett and Troendle, 1975). Early concepts of stormflow runoff suggested that the only mechanism capable of producing quick-responding peak discharge was surface runoff. The VSAC suggests that the two mechanisms that are primarily responsible for the rapid flow response are an expanding source (saturated) area that contributes flow directly to a channel and a rapid subsurface flow response from upland to lowland areas (Fig. 5.11).

A stream channel and the riparian corridor immediately adjacent to the channel respond most quickly to a rainfall or snowmelt event (see Chapter 6). As rain falls and a snowpack melts, the wetter areas and areas of shallow soil become saturated, after which this saturated zone expands both upslope and upstream. As a result, the area of a watershed contributing directly to the channel becomes increasingly larger with the duration of the storm. This source area slowly shrinks once the rainfall or snowmelt stops.

Water stored in the soil upslope from the channel system contributes to flow downstream by displacement and direct flow out of saturated zones near the channel. Midslope and low areas can respond quickly due to the displacement of upslope water into the saturated zone. Ridgetop areas might contribute little to stormflow. Much of the infiltrated water can be stored or the pathway can be long enough to delay the flow until long after the storm event has ended.

Factors Affecting Stormflow Response

The factors that determine the magnitude of stormflow volume and peak flow can be separated into those that are fixed and those that vary with time. The factors that are fixed in time and have a pronounced influence on stormflow response include the size and shape of the watershed, the steepness of the hillslopes, the character of the stream channels of the watershed, and the presence of lakes, wetlands, or other water bodies within the watershed boundary. Typically, the larger the watershed, the greater the volume and peak of streamflow from rainfall or snowmelt-runoff events; however, storage must be factored into this assumption.

Watershed shape affects how quickly surface and subsurface flows reach the outlet of a watershed. Higher peak flows will tend to occur with a round-shaped watershed, concentrating surface runoff more quickly at the outlet than an elongated watershed. Likewise, the steeper the hillslopes and channel gradients, the quicker the response and the higher the peak stormflows.

Drainage density, the sum of all stream channel lengths divided by the watershed area, also affects the rapidity with which water flows to the outlet. A sharply rising hydrograph and a high peak that is referred to as a flashy stream is typically associated with higher drainage densities. As the percentage of watershed area in lakes, ponds, or wetlands increases, a greater attenuation or flattening of the stormflow hydrograph occurs. Such water bodies can increase the travel time through the watershed to the outlet. Wetlands and lakes detain (slow down travel time through the process of reservoir routing) and retain (store) water flowing into them.

The factors affecting stormflow response that vary with time can be separated into the precipitation inputs and watershed conditions. The magnitude of rainfall or snowmelt runoff affects the magnitude of stormflow response. The intensity and duration are especially important for rainstorms. As a general rule, the higher the intensity and the longer the duration of a rainstorm, the higher the magnitude of peak flow. The areal distribution of a rainstorm and the movement (tracking) of the event affect the peak and volume of stormflow. A storm that moves from the upper reaches of a watershed to the outlet will tend to concentrate flow at the outlet, whereas one that moves upstream will tend to spread out the flow response over time. Such a response is due to the timing of drainage and the routing of flow from upland and from downstream parts of a watershed.

The length of time between rainfall or snowmelt-runoff events affects the *antecedent conditions* of a watershed, which refers to the relative moisture storage status of a watershed at a point in time. If a watershed has experienced rainfall or snowmelt recently, it is likely "primed" to respond quicker and with a greater volume of streamflow because there is less available storage than the one that has not had precipitation for weeks.

Watershed conditions that can vary and influence stormflow response include vegetation type and extent of cover, soil surface conditions, and a variety of human-caused changes such as roads, reservoirs, drainage systems, waterways, and stream channel alterations. These human-caused changes can be manipulated by people to achieve desired hydrologic objectives or to become altered as part of development or other management activities. All these factors exert some influence on stormflow response. However, it is difficult to separate and quantify the contributions of individual components or factors. Computer simulation models and other hydrologic methods (see Chapters 6 and 16) have been developed to quantify the various watershed and meteorological factors affecting stormflow response.

Summary and Learning Points

The runoff response from a watershed due to rainfall or snowmelt-runoff events is the integrated effect of many factors. Some of these factors are affected directly by human activities on the watershed, whereas others are not. To this point in the book, many of the precipitation and watershed characteristics that affect infiltration, percolation, recharge, and runoff type have been discussed. By now, therefore, you should be able to:

1. Explain how the following affect infiltration rates:
   - soil moisture content
   - hydraulic conductivity of the soil
   - soil surface conditions
   - presence of impeding (water repellent) layers in the soil profile.
2. Explain how land-use activities affect the infiltration capacities of a soil through each of the above.
3. Discuss how changes in infiltration capacities can result in different flow pathways through the watershed.
4. Illustrate and discuss the different pathways and mechanisms of flow that result in percolation, recharge, surface runoff, and a stormflow hydrograph for a forested watershed with deep soils in comparison to an agricultural or urban watershed. Explain how the major pathways of flow differ in each case.
5. Explain how groundwater recharge occurs in different landscape settings.
6. Explain translatory flow and variable source area concept.

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CHAPTER 6 *Streamflow Measurement and Analysis*

INTRODUCTION

Streamflow is the primary mechanism by which water moves from upland watersheds to ocean basins. Streamflow is the principal source of water for municipal water supplies, irrigated agricultural production, industrial operations, and recreation. Sediments, nutrients, heavy metals, and other pollutants are also transported downstream in streamflow. Flooding occurs when streamflow discharge exceeds the capacity of the channel. Therefore, streamflow measurements or methods for predicting streamflow characteristics are needed for various purposes. This chapter discusses the measurement of streamflow, methods and models used for estimating streamflow characteristics, and methods of streamflow-frequency analysis.

MEASUREMENT OF STREAMFLOW

Streamflow discharge, the quantity of flow per unit of time, is perhaps the most important hydrologic information needed by a hydrologist and watershed manager. Peak-flow data are needed in planning for flood control and designing engineering structures including bridges and road culverts. Streamflow data during low-flow periods are required to estimate the dependability of water supplies and for assessing water quality conditions for aquatic organisms (see Chapter 11). Total streamflow and its variation must be known for design purposes such as downstream reservoir storage.

The *stage* or height of water in a stream channel can be measured with a staff gauge or water-level recorder at some location on a stream reach. The problem then becomes converting a measurement of the stage of a stream to discharge of the stream. This problem

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is solved by either *stream gauging* or installing precalibrated structures such as *flumes* or *weirs* in the stream channel.

Measuring Discharge

A simple way of estimating discharge is to observe the time it takes for a floating object that is tossed into the stream to travel a specified distance. A measurement of the cross section of the stream should be made simultaneously and the two values are then multiplied together:

$$Q = VA \tag{6.1}$$

where Q is the discharge (m^3/s); V is the velocity (m/s); and A is the cross section (m^2).

This method of estimating discharge is not accurate, however, because velocity of a stream varies from point to point with depth and width over the cross section of the stream. The velocity at the surface is greater than the mean velocity of the stream. Actual velocity is generally assumed to be about 80–85% of surface velocity.

The velocity profile can be estimated by dividing the cross section of a stream into vertical sections and measuring the mean velocity of the stream at each section (Fig. 6.1).

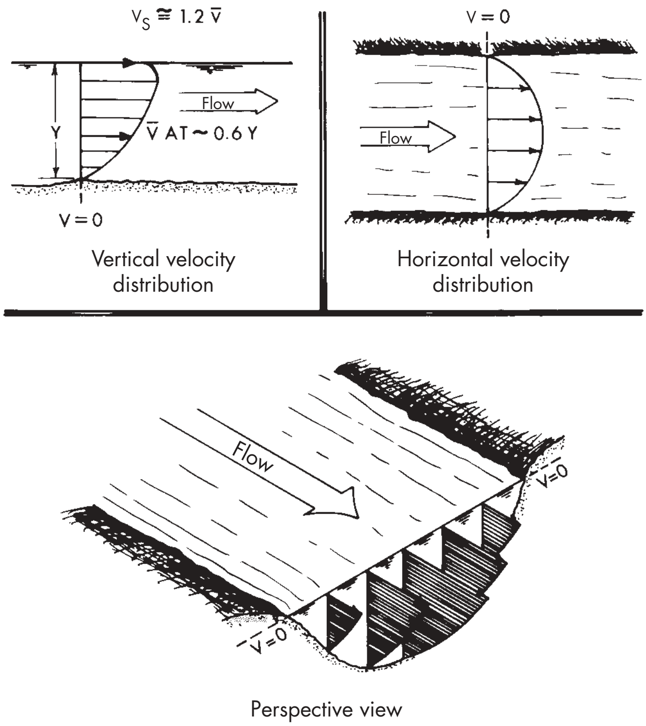

FIGURE 6.1. Measurements of stream channel cross sections and velocities needed to estimate the mean velocity of a stream

The area of each section can be determined and the average discharge of the entire stream computed as the sum of the product of area and velocity of each section as follows:

$$Q = \sum_{1}^{n} A_i V_i \tag{6.2}$$

where n is the number of sections.

The greater the number of sections, the closer the approximation to the true value. For practical purposes, however, between 10 and 20 sections are commonly used depending on the channel width. The actual number of sections depends on the channel configuration and the rate of change in the stage with discharge. Depth and velocity should not vary greatly between points of measurement. Importantly, all of the measurements should be completed before the stage changes too much.

The velocity and depth of vertical sections can be measured by wading into the stream or from cable car, boat, or bridge. Velocity of the stream is usually measured by a current meter with the following general rules:

- Two measurements are made for each section at 20% and 80% of the total depth and then averaged for depths >0.5 m.
- One measurement is made at 60% of the depth for depths <0.5 m.
- A current meter smaller than a standard current meter, such as a pygmy current meter, is used to measure velocity of shallow streams less than about 0.5 m deep.

The most critical aspect of stream gauging is the selection of a control section, that is, a section of the stream for which a *rating curve* (Fig. 6.2) is to be developed. This section should be stable, have a sufficient depth for obtaining velocity measurements at the lowest of streamflows, and be located in a straight reach without turbulent flow.

Further information concerning measurement of streamflow and analysis of streamflow data can be obtained from the US Geological Survey (accessed July 20, 2011;

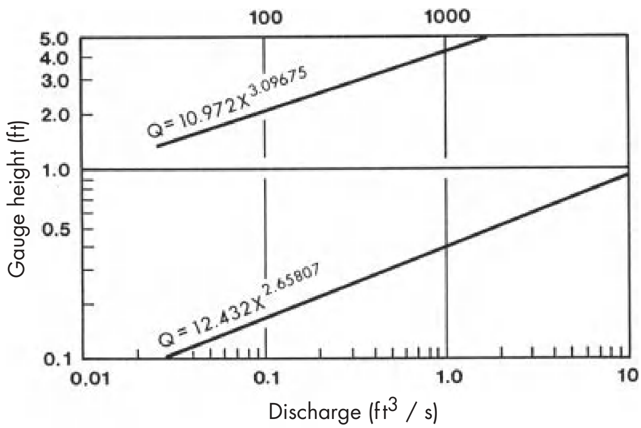

FIGURE 6.2. Example of a rating curve of streamflow discharge in relation to depth measurements (from Brown, 1969). A logarithmic relationship is shown in the figure because of the range of measurements obtained

http://ga.water.usgs.gov/edu/measureflow.html) or Oregon State University (accessed July 20, 2011; http://streamflow.engr.oregonstate.edu/manipulation/index.htm).

In the United States, the United States Geological Survey (USGS) measures streamflow discharge directly with an acoustic Doppler current profiler (ADCP) system (Mueller and Wagner, 2009; accessed July 20, 2011; http://pubs.water.usgs.gov/tm3a22). Acoustic energy is transmitted at a known frequency and the measured backscatter of acoustic energy is used to compute the velocity of water particles. Discharge can be determined by deploying the ADCP system by towing or remotely controlling a moving boat to determine the velocity in the water column from one streambank to the other. This system determines instantaneous streamflow discharge directly in the channel.

Precalibrated Structures for Streamflow Measurement

Precalibrated structures are often used on small watersheds, usually less than 800 ha in size because of their convenience and accuracy. The most common precalibrated structures are *weirs* and *flumes*. Because of their greater accuracy, weirs are often used for gauging small watersheds, particularly those with low flows. Flumes are preferred where sediment-laden streamflows are common.

Weirs

A weir is a small wall or dam across a stream that forces water to flow through a notch with a designed geometry. The notch can be V-shaped, rectangular, or trapezoidal (Fig. 6.3). Water is impounded upstream of the dam, forming a *stilling basin* that is connected to a water

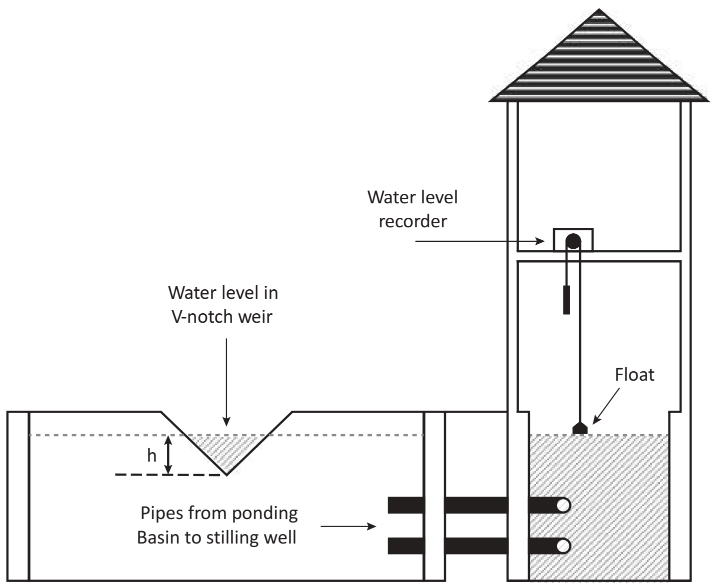

FIGURE 6.3. A schematic illustration of a weir with a V-shaped notch on the weir blade

level recorder. A gaugehouse or other type of shelter is provided to protect the recorder. The cutoff wall (dam) used to divert water through the notch is seated into bedrock or other impermeable material where possible so that no water can flow under or around it. The stilling basin is sometimes constructed as a watertight box where leakage is apt to occur.

The notch edge or surface over which the water flows is called the *crest*. Weirs can be either sharp crested or broad crested. A sharp-crested weir has a blade with a sharp upstream edge so that the passing water touches only a thin edge and clears the rest of the crest. This design provides accurate measurements at low flows. A broad-crested weir has a flat or broad surface over which the discharge flows and is used where sensitivity to low flows is not critical.

The rectangular weir has vertical sides and horizontal crests. Whereas its major advantage is its capacity to handle high flows, the rectangular weir does not provide precise measurement of low flows. The trapezoidal weir is similar to the rectangular weir, but it has a smaller capacity for the same crest length. The discharge is approximately the sum of discharges from the rectangular and triangular sections. Sharp-crested V-notch or triangular weirs are often used where accurate measurements of low flows are important. The V-notch weirs can have a high rectangular section to accommodate infrequent high flows.

Flumes

A flume is an artificial open channel built to contain flow within a designed cross section and length. There is no impoundment, but the height of water in the flume is measured with a stilling well. Several types of flumes have been used on small watersheds. Figure 6.4 shows a trapezoidal broad-crested flume that is used to measured open channel flow (Robinson and Chamberlain, 1960). HS-, H-, and HL-type flumes developed and rated by the U.S. Natural Resources Conservation Service (NRCS) have converging vertical sidewalls cut

FIGURE 6.4. A broad-crested trapezoidal flume for measuring open channel streamflow

back on a slope at the outlet to give them a trapezoidal projection. These flumes have been used largely to measure intermittent runoff events.

The Venturi flume is rectangular, trapezoidal, triangular, or any other regular shape with a gradually contracting section leading to a constricted throat and an expanding section immediately downstream. The floor of the Venturi flume is situated at the same grade as the stream channel. Stilling wells for measuring the head are situated at the entrance and the throat. The difference in head at the two wells is related to discharge. This type of flume is used widely to measure irrigation water. The Parshall flume is a modification of the Venturi flume that measures water in open conduits. It is also frequently used for measuring irrigation water.

The San Dimas flume measures debris-laden flows in mountain streams. It is rectangular, has a sloping floor (3% gradient), and functions as a broad-crested weir except that the contraction is from the sides rather than the bottom. Therefore, there is no barrier to cause sediment deposition. Rapid flow keeps the flume scoured clean.

The need to measure both high- and low-volume sediment-laden flows accurately prompted the development of an improved design of a supercritical flume (Smith et al., 1981). This flume was used initially at the Santa Rita Experimental Range and the nearby Walnut Gulch Experimental Watersheds in southeastern Arizona. The flume was intended for use in small channels with flows of generally less than 4 m^3/s. However, the flume size is not limited as long as certain proportions are maintained.

Considerations for Using Precalibrated Structures

The type of weir or flume to use in a specified situation depends largely on the estimated magnitudes of the maximum and minimum flows, the accuracy needed in determining total discharge for high flows and low flows, the amount and type of sediments or other debris expected, the stream channel gradient and cross section, the underlying material, the accessibility to the site, the anticipated length of time of streamflow monitoring, and funding available for establishing the structure. Weirs are generally more accurate than flumes at low flows, whereas flumes are often preferred when high volumes of sediment or other debris are transported in the flow.

Flumes and weirs have been used together in tandem. Under high flows, the discharge flows through the flume and over the weir that is located immediately downstream. Under low flows, there is not a sufficient jump to clear the downstream weir. In these cases, the water trickles into the impoundment above the weir, and accurate measurements of low flow can be obtained. However, such configurations are costly and can be justified usually only for experimental purposes.

Monitoring Streamflow in the United States

The USGS collects and distributes information on the water resources of the United States to interested people. Included in this information are water data, related publications and maps, and statistics on the status of recent water-related projects. Streamflow measured on a network of gauges operated by the USGS on a state-by-state basis is also provided in real time; these data can be accessed by a link to the USGS website (http://water.usgs.gov).

The US Army Corps of Engineers and US Department of the Interior (USDI) Bureau of Reclamation monitor the flow of water from selected watersheds and river basins with summaries of the streamflow measurements obtained available on their respective websites. Many states, counties, and municipalities also monitor streamflows on selected basins to determine the availability of water resources.

Streamflow is monitored on experimental watersheds maintained by the U.S. Forest Service, USDA Agricultural Research Service, and other agencies throughout the country. Summaries of these measurements facilitate evaluations of possible changes attributed to climatic variability, effects of watershed management practices including the construction of roads, trails, and other corridors, and impacts of wildfire and pest outbreaks on streamflow regimes. These evaluations are a basis for managing wildland, agricultural, and urban watersheds and determining the impacts of changing land use and possible climatic change on these watersheds (discussed later in the book).

Obtaining Discharge Information on Ungauged Streams

Streamflow information is often needed where there is no gauge to measure streamflow. Such is the case on small streams in many rural areas. Even on large rivers there can be long reaches without stream gauges. An estimation of streamflow discharge can be needed for a variety of purposes, but frequently hydrologists and watershed managers are concerned with the magnitude of flood flows. Approximations of streamflow can sometimes be made by extrapolating information from a similar basin that is instrumented with a gauge. This extrapolation must be done cautiously by an experienced analyst because these estimates are never as good as direct measurements of streamflow.

Two commonly used methods for estimating stream discharge at known stages (depths) of flow where no stream gauge exists are the Manning and the Chezy equations.

The *Manning equation* is:

$$V = \frac{1.49}{n} R_h^{2/3} S^{1/2} \tag{6.3}$$

where V is the average velocity in the stream cross section (ft/s); R_h is the hydraulic radius (ft), which is equal to A/WP where A is the cross-sectional area of flow (ft^2) and WP is the wetted perimeter (ft); s is the energy slope as approximated by the water surface slope (ft/ft); and n is a roughness coefficient.

The *Chezy equation* is:

$$V = C\sqrt{R_h S} \tag{6.4}$$

where C is the Chezy roughness coefficient.

Equations 6.4 and 6.5 are similar. The relationship between the roughness coefficients is:

$$C = \frac{1.49}{n} R_h^{1/6} \tag{6.5}$$

The two equations are used in a similar fashion. The hydraulic radius and water surface slope are obtained from cross-sectional and bed-slope data in the field (Fig. 6.5). Manning's roughness coefficient is estimated (Table 6.1) and the average discharge, Q, is calculated by multiplying the velocity by the cross-sectional area (A in Fig. 6.5).

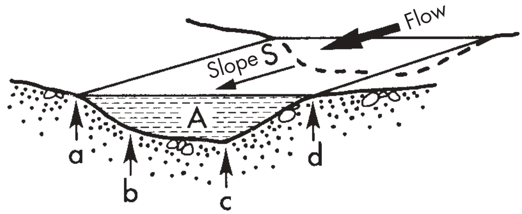

FIGURE 6.5. Stream channel section showing the slope or gradient of the streambed, wetted perimeter *WP* (a, b, c, d), and cross-sectional area A

Manning and Chezy equations are most often used to estimate a previous peak streamflow. High watermarks located after a stormflow event has ended can be used to estimate the depth of flow. Sometimes, this estimate is obtained by measuring the height of debris caught along the stream channel (Fig. 6.6) or watermarks on structures. The estimate of high water should be obtained for a reach of stream channel where the cross section of the flow of interest can be measured with reasonable accuracy. The slope of the water surface can be approximated by the slope of the channel along the reach. The wetted perimeter can be measured by laying a tape on the channel bottom and sides between the high watermarks. The cross-sectional area should be measured by summing several segments (Fig. 6.5 and Fig. 6.6).

METHODS FOR ESTIMATING STREAMFLOW CHARACTERISTICS

Analytical methods are available for estimating streamflow characteristics such as peak stormflows, stormflow volumes, and low flows when this information has not been, or cannot be, obtained by on-site monitoring. We often need some way of estimating the streamflow response of a watershed to a meteorologic event or climatic condition that has not before been observed at a given location. Furthermore, planners and managers might be interested in estimating how streamflow might change in response to changes in land use on a watershed.

Methods of estimating streamflow range from the transfer of streamflow information from gauged to ungauged watersheds, to applying simple predictive regression equations, to more generally applied methods such as a unit hydrograph (UHG) analysis, and to the use of comprehensive computer simulation methods. Selecting the appropriate method for a specified estimation requires consideration of the following:

- type and accuracy of information required;
- available data;
- physical and biological characteristics of the watershed in question;
- technical capabilities of the individual performing the study; and
- time and economic constraints.

Table 6.1. Examples of Manning's roughness coefficient *n*

Type of channel	Minimum	Average	Maximum
A. Excavated or dredged			
1. Earth, straight, clean	0.016	0.018	0.020
2. Gravel, uniform, clean	0.022	0.027	0.033
3. Earth, winding, sluggish, grass, some weeds	0.025	0.030	0.033
4. Dragline excavated, light brush on banks	0.035	0.050	0.060
5. Channels not maintained, weeds/brush not cut	0.050	0.080	0.120
B. Natural streams			
1. Minor streams with width at flood stage <30 m			
a. Streams on plains			
• Clean, straight, full stage no pools	0.025	0.030	0.033
• Clean, winding, some pools and bars	0.033	0.040	0.045
• Same as above but some weeds and stones	0.035	0.045	0.050
• Sluggish reaches, weedy, deep pools	0.050	0.070	0.080
• Very weedy, deep pools or floodways with heavy stand of timber and underbrush	0.075	0.100	0.150
b. Mountain streams, no vegetation in channel, banks usually steep, brush along banks submerged at high stages			
• Bottom consists of gravels, cobbles and few boulders	0.030	0.040	0.050
• Bottom consists of large boulders and some large organic debris, sinuous flow	0.050	0.070	0.100
2. Floodplains			
a. Pasture, no brush, short grass	0.025	0.030	0.035
b. Pasture, no brush, tall grass	0.030	0.035	0.050
c. Cultivated areas, no crop	0.020	0.030	0.040
d. Cultivated areas, mature row crops	0.025	0.035	0.045
e. Scattered brush, heavy weeds	0.035	0.050	0.070
f. Medium to dense brush in winter	0.045	0.070	0.110
g. Medium to dense brush in summer	0.070	0.100	0.160
h. Dense willows, summer, straight channel	0.110	0.150	0.200
i. Dense stand of timber, few downed trees, little undergrowth, flood stage below branches	0.080	0.100	0.120
j. Same as above but flood stage above branches	0.100	0.120	0.160
3. Major streams with width at flood stage > 30 m			
a. Streams on plains			
• Sand channels	0.025	0.035	0.045
• Boulder channels	0.028	0.040	0.045
• Vegetation-lined channels at flood stage	0.045		0.120
b. Mountain streams			
• Cobbly bottoms, no debris dams	0.028	0.035	0.040
• Cobbly bottoms with debris dams	0.032		0.060
• Large boulders, debris dams in channel	0.050		0.100

Source: Adapted from Gray (1973) and Van Haveren (1986).

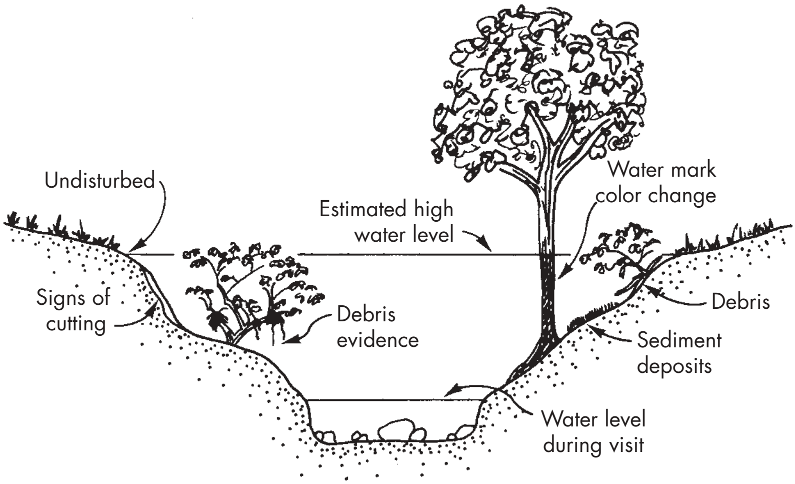

FIGURE 6.6. Evidence of high watermarks in a stream for estimating the depth of flow

Direct Transfer of Streamflow Information

Streamflow information can be transferred from a gauged to an ungauged watershed if the two watersheds are deemed to be hydrologically similar. The hydrologic similarity of the watersheds depends largely on the following criteria:

1. Both of the watersheds should be located within the same meteorological regime.
2. Soils, geology, topographic relief, watershed shape, drainage density, and type and extent of vegetative cover and land use should be similar.
3. The drainage areas of the watersheds should be about the same size and preferably within an order of magnitude.

Direct transfer of streamflow information is generally employed to obtain approximations of streamflow characteristics in one of the two ways. The entire long-term (historical) streamflow record from the gauged watershed can be transferred to the ungauged watershed with adjustments made for differences in the drainage areas. For example, the observed streamflow record could be multiplied by the ratio of ungauged to gauged drainage areas. The second approach is to transfer only specific streamflow characteristics such as the 100-year return period of the annual flood peak or the minimum 7-day flow, once again adjusting for drainage area differences. The entire flood-frequency curve for an ungauged watershed can be estimated by adjusting a frequency curve developed for a hydrologically similar gauged watershed.

The *direct-transfer method* is quick and easy to use and normally employed for rough approximations. However, it is applicable only if the assumption of hydrologic similarity is met. If a significant amount of adjustment is needed to transfer hydrologic information from a gauged to an ungauged watershed, there can be little confidence in the result.

TABLE 6.2. Values of runoff coefficients (C) for the rational method

Soil type	Cultivated	Pasture	Woodlands
With above-average infiltration rates; usually sandy or gravelly	0.20	0.15	0.10
With average infiltration rates; no clay pans; loams and similar soils	0.40	0.35	0.30
With below-average infiltration rates; heavy clay soils or soils with a clay pan near the surface; shallow soils above impervious rock	0.50	0.45	0.40

Source: Adapted from ASCE (1969) and Dunne and Leopold (1978).

Estimating Peak Discharge

The *rational method* is commonly applied for estimating the peak stormflow discharge from rainfall events occurring on small watersheds having relatively uniform land-use conditions. The underlying basis of this method is that the maximum rate of runoff occurs when the entire watershed area contributes to stormflow at the outlet. The estimates of peak discharge are valid only for storms in which the rainfall period is at least as long as the watershed *time of concentration*. Time of concentration (T_c) is the time required for the entire watershed to contribute streamflow at the outlet or, more specifically, the time it takes for water to travel from the most distant point on the watershed to the watershed outlet.

Peak storm discharge from homogeneous watersheds smaller than 1000 ha is estimated from the following equation:

$$Q_p = \frac{C P_g A}{K_m} \tag{6.6}$$

where Q_p is the peak stormflow discharges (m^3/s); C is the runoff constant (see Table 6.2); P_g is the rainfall intensity (mm/h) of a storm with a duration at least equal to the time of concentration (T_c) on the watershed; A is the area of the watershed (ha); and K_m is a constant, 360 for metric units (1 for English units).

The assumption that rainfall intensity is uniform over the entire watershed for a period equal to the time of concentration is seldom met under natural conditions and would apply only for small watersheds. The method is commonly used for designing stormwater systems in urban landscapes and determining the size of culverts below small catchments (Box 6.1). For most applications, therefore, regional rainfall intensity and duration values that are associated with an acceptable risk or frequency are used. In Box 6.1, the 25-year recurrence-interval rainfall for 1-hour duration (T_c for the watershed) in southeastern Minnesota was 2.4 in. (Huff and Angel, 1992).

Several "simple" methods similar to the rational method have been developed in many parts of the world (see Gray, 1973 for examples). Importantly, all such methods should be used only for quick estimates of peak flows from small watersheds where assumptions for use and data requirements are met. Another simple method for estimating peak flows over time is the USGS crest gauge. The crest gauge is measured once per year at a given site for the maximum stage height. Several years of data are then used to infer recurrence interval.

Box 6.1

Example of Applying the Rational Method to Determine the Size of a Culvert

A culvert system is to be designed to pass a 25-year recurrence-interval rainfall event for a 90 ac woodland watershed in southeastern Minnesota (Midwestern USA) based on the following information:

- Soils are sandy loam with high infiltration rates; from Table 6.2, a runoff coefficient C of 0.10 was chosen.
- Time of concentration for the watershed was estimated to be 60 minutes.

From Bulletin 71, Rainfall Frequency Atlas of the Midwest (Huff and Angel, 1992 and http://www.isws.illinois.edu/pubdoc/B/ISWSB-71.pdf, accessed July 21, 2011), a 25-year recurrence-interval rainfall for a duration of 1 hour is 2.4 in. in southeastern Minnesota.

Applying the Rational Method, the peak discharge to design the culvert system would be as follows:

$Q = C\,P_g\,A$
(0.10) (2.4 in/h) (90 ac)
21.6 cfs

Stormflow Response

The stormflow response of a watershed can be characterized by separating stormflow from baseflow (Box 6.2). Several methods are available to estimate stormflow volumes and stormflow hydrographs ranging from regionally developed stormflow response factors to the more generalized UHG and *curve number* (CN) methods.

Stormflow Response Factor

Simpler approaches can also be used such applying the following hydrologic response relationship (Woodruff and Hewlett, 1970):

$$R_s = \left(\frac{\text{annual stormflow}}{\text{annual precipitation}}\right)100 \qquad \textbf{(6.7)}$$

where R_s is the hydrologic response (%).

This hydrologic response relationship provides an indication of the stormflow response or flashiness of a watershed to rainstorms. The hydrologic response for individual storms can vary from less than 1% to more than 75% depending largely on the antecedent moisture conditions (AMCs) of the watershed.

Box 6.2

Separating Baseflow from Total Streamflow Resulting from a Storm Eevent

Baseflow (delayed flow) must be separated from total streamflow for several single-event methods of stormflow analysis. The stormflow volume is that portion of the hydrograph above baseflow and is sometimes called direct runoff or quickflow. There is no universal standard of separating baseflow because flow pathways through a watershed cannot be directly related to the hydrograph that represents the integrated response of all flow pathways. This does not present a problem for most flood analyses, however, because the baseflow contribution is typically less than 10% of the stormflow. Therefore, efforts to devise elaborate baseflow-separation routines are usually not warranted. The following steps are recommended:

1. Graphically separate baseflow from stormflow for several storm hydrographs, such as with methods I and II in the figure below. Once one method has been adopted, it should be used for all analyses.
2. After examining several storms, determine if there is a consistent relationship that can be expressed as follows:
   - Draw a straight line from the beginning point of hydrograph rise to a point on the recession limb defined by *N* days after the peak, where *N* is A^c; *A* is the watershed area in square miles; *c* is a coefficient (typically a value of 0.2 is used; see Linsley et al., 1982).
   - Determine if the separation line I yields a consistent rate in terms of cfs per square mile per hour. Hewlett and Hibbert (1967) found that 0.05 cfs/mi^2/h was satisfactory for watersheds in the southeastern USA.

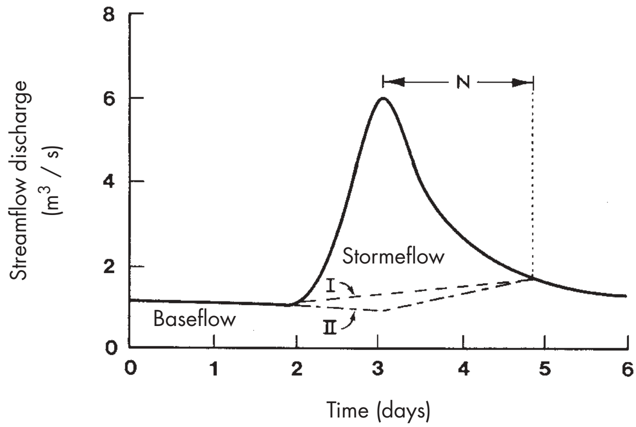

Figure for Box 6.2. Methods of separating baseflow from stormflow; I and II are different approaches.

The stormflow response of a watershed to climatic variability, management practices, and incidences of fire, insects, and disease can be characterized by calculating the average ratio of *stormflow volume* to the amount of precipitation of the period of the stormflow. The precipitation can be rainfall, melting of snowfall, or a combination both. Stormflow volume is determined by separating *baseflow* from the total streamflow resulting from the storm event. By comparing stormflow response factors for watersheds with different vegetative cover and land uses within the same climatic region, the flood-producing potentials for different areas can be estimated. Knowledge of the stormflow response factor can also be used to develop estimates of peak discharge where a consistent relationship between stormflow volume and peak exists.

A constraint in applying this method is that both precipitation and stormflow data must be available. Determining the stormflow response factor is also dependent on the AMC of the watershed. However, the lack of this information does not necessarily preclude a regional analysis if regression models can be used to predict response factors based on measurable watershed characteristics such as size, slope, vegetative cover, and land use.

Analyses of stormflow on small watersheds in Georgia by Hewlett and Moore (1976) resulted in equations of the form:

$$Q_s = 0.22R\,(SD)P^2 \tag{6.8}$$

where Q_s is the stormflow (in.); SD is the sine-day factor, which is equal to sin [360(day number/365)] + 2, where day 0 is November 21; and P is the total storm precipitation on the watershed (in.); and

$$R = \sum_{1}^{n} (Q_s/P)/n \tag{6.9}$$

for n observations and P is greater than or equal to 1 in. The sine-day factor approximates AMCs as a seasonal coefficient for the region.

Comparisons of stormflow volumes and peaks (Q_p) resulted in the following equation:

$$Q_p = 25\,R(SD)^{0.5}P^{2.5} \tag{6.10}$$

where Q_p is the peak discharge above baseflow (cfs/mi^2).

An application of the stormflow response factor method (Equation 6.8) is shown in Box 6.3. Simple relationships like the response factors are useful in hydrology and watershed managers but should only be applied to watersheds within the region in which they were developed and tested.

Unit Hydrograph

One of the more widely used methods of analyzing stormflow is the UHG method. A UHG is the hydrograph of stormflow resulting from 1 unit (1 mm or 1 in depending on units used) of effective precipitation occurring at a uniform rate for a specified time period and areal distribution on a watershed. It represents stormflow response as shown by the shape of the hydrograph for the watershed. The *effective precipitation* is the amount of rainfall or snowmelt in excess of watershed storage requirements, groundwater contributions, and evaporative losses. In essence, it is the portion of total precipitation that ends up as stormflow. Therefore, the volume of effective rainfall equals the volume of stormflow.

Box 6.3

Applying the Response Factor Method to Estimate Storage Requirements

The amount of storage required to hold the total discharge from a 24-hour, 50-year return period rainstorm (7.5 in.) on a 3000-ac watershed is estimated. The watershed has a 70% cover of old-growth forest ($R = 0.10$) and 30% of the area is pasture and cultivated agricultural land ($R = 0.18$). Assume $SD = 2.99$ (February storm).

Forested

$Q_s = (0.22)\ (0.10)\ (2.99)\ (7.5)^2 = 3.70$ area in.
volume = (3.7 area in.) (2100 ac) (1 ft/12 in.)
647.5 ac-ft

Pasture and cultivated

$Q_s = (0.22)\ (0.18)\ (2.99)\ (7.5)^2 = 6.66$ area in.
volume = (6.66 area in.) (900 ac) (1 ft/12 in.)
499.50 ac-ft

Total storage required is 1147 ac-ft.

The UHG method is a *black-box method* that relates stormflow output to a specified duration of precipitation input. No attempt is made to simulate the hydrologic processes involved in the flow of water through the watershed. The UHG concept provides the basis for several hydrologic models of greater complexity and wider application than simplified formulas such as the rational method.

Development of a Unit Hydrograph. The UHG concept is best understood by examining the method of developing a UHG from an isolated storm (Fig. 6.7). Records of the watershed are examined initially for single-peaked, isolated streamflow hydrographs that result from short-duration, uniformly distributed rainfall or snowmelt hyetographs of relatively uniform maximum intensity. Once a hyetograph–hydrograph pair is selected, dividing streamflow by the watershed area converts the scale of the hydrograph to mm or in of depth. Separating the more uniform baseflow from the rapidly changing stormflow component determines the total stormflow depth from the hydrograph (Fig. 6.7). Each ordinate of the stormflow hydrograph is then divided by the total stormflow depth, resulting in a normalized hydrograph of unit value under the curve (Fig. 6.7).

The area-weighted hourly precipitation distribution that caused the stormflow hydrograph is the next factor to be determined (Fig. 6.7). In this case, the *effective rainfall* is defined as being equal to stormflow. Therefore, the total interception, storage, and deep seepage loss is estimated as the difference between total rainfall or snowmelt and total stormflow (Fig. 6.7). The magnitude of these losses is largely a function of the AMCs.

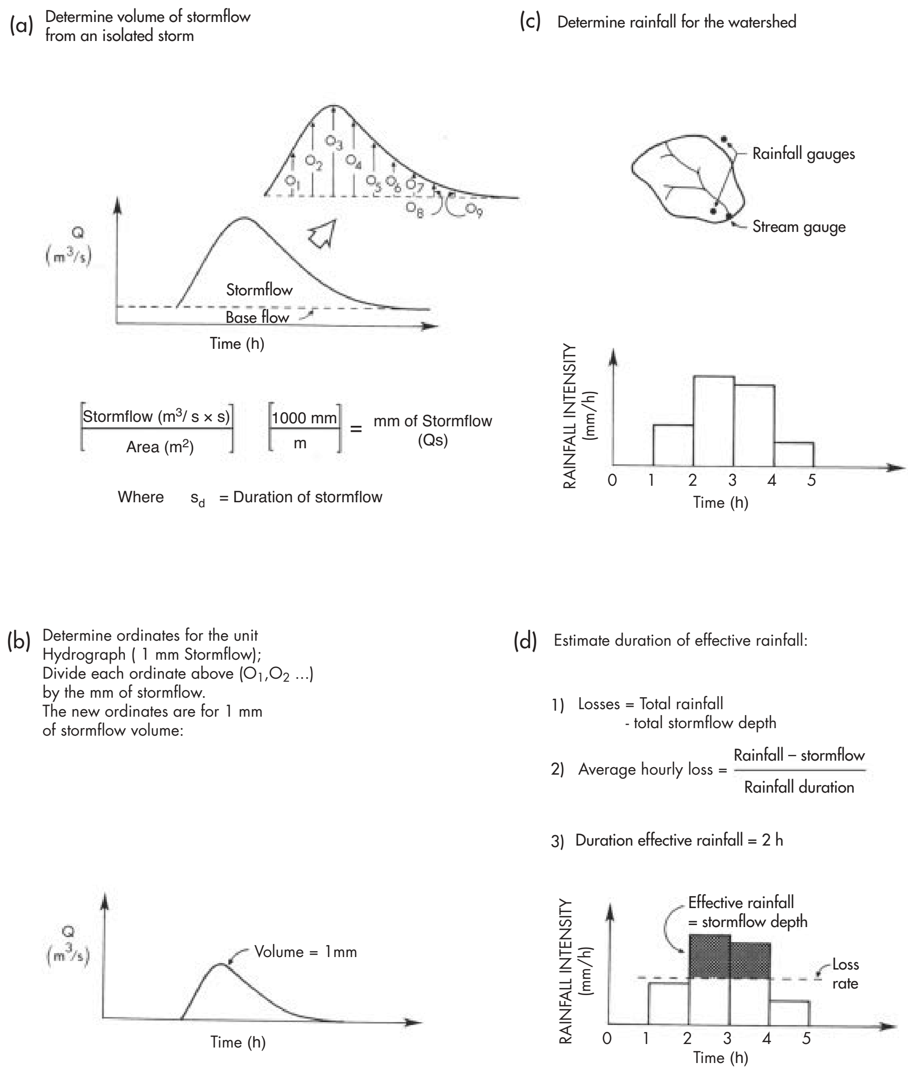

FIGURE 6.7. Development of a UHG for an isolated storm

In Fig. 6.7, a uniform-loss rate was assumed, but a diminishing-loss curve can be used if available. The effective rainfall in Figure 6.7d was uniform during the 2-hour period bracketed by the loss curve. The duration of effective rainfall characterizes the UHG not the duration of the UHG itself. For example, a UHG developed from an effective rainfall of 3/4-hour duration is a 3/4-hour UHG, that of 1-hour duration is a 1-hour UHG, and that of 6-hour duration is a 6-hour UHG. UHGs can also be developed from multipeaked stormflow hydrographs by successive approximations. These complex storms are best analyzed with computer assistance.

The normalized shape of the UHG is characteristic of a watershed for a specified intensity of effective rainfall and, therefore, is an index of stormflow for the watershed.

Table 6.3. Application of a 1-h UHG to a storm of 2 h of effective precipitation

Time (h)	1-h UHG ordinates (m^3/s)	Effective rainfall (mm)	Stormflow: Time 1 (m^3/s)	Stormflow: Time 2 (m^3/s)	Stormflow: Subtotal (m^3/s)	Base flow (m^3/s)	Total discharge (m^3/s)
0	0.00	0	0.00	0.00	0.00	1.2	1.2
1	0.05	20	1.00	0.00	1.00	1.2	2.2
2	0.50	30	10.00	1.50	11.50	1.2	12.7
3	1.00	0	20.00	15.00	35.00	1.2	36.2
4	0.75	0	15.00	30.00	45.00	1.2	46.2
5	0.50	0	10.00	22.50	32.50	1.2	33.7
6	0.25	0	5.00	15.00	20.00	1.2	21.2
7	0.00	0	0.00	7.50	7.50	1.2	8.7
8	0.00	0	0.00	0.00	0.00	1.2	1.2

It, therefore, represents the integrated response of that area to a rainfall input. The UHG method assumes that the effective rainfall and loss rates are relatively uniform over the entire watershed area. The watershed characteristics that affect stormflow response must also remain constant from the time that the UHG is developed until the time it is applied.

Application of a Unit Hydrograph. The rainfall quantity and distribution over time from a design storm is obtained initially for the watershed. Estimated loss rates are then subtracted from total rainfall to obtain the quantity, distribution, and duration of effective rainfall. The rainfall duration is divided by the selected UHG duration to obtain the number of periods to be added up for the total design storm. The ordinates of the selected UHG are then multiplied by the quantity of effective rainfall for each period. In Table 6.3, for example, 20 mm of effective rainfall yields a stormflow hydrograph with ordinates 20 times those of the corresponding UHG for the first period and 30 mm of effective rainfall is 30 times the UHG but delayed for the second period. The calculated stormflows plus any baseflow are summed for each period to obtain the total stormflow hydrograph.

The assumption of linearity is not always valid. As effective rainfall increases, the magnitude of the peak actually can increase more than the proportional increase in the rainfall amount. A consequence of ignoring a nonlinear response can be to underestimate the magnitude of peak discharge for large storm events. If a nonlinear response is suspected, two or more UHGs should be developed from observed hydrographs resulting from substantially different rainfall amounts. The appropriate UHG would then be applied only to precipitation amounts similar to those used to develop the UHG in the first place.

The UHG developed from a specific duration of effective rainfall can be applied only to effective rainfall that occurred over the same duration. That is, a 6-hour UHG cannot be applied directly to analyze a storm with an effective rainfall that occurred over 2 hours. The duration of a UHG must be changed in such cases. The most common method of converting a UHG from one duration to another by applying the *S-curve method* is shown in the following figure. Once a UHG of a specified duration is developed, for example, for 4 hours, an S-curve can be constructed by adding the UHGs together with each lagged by its duration. Summation of the lagged UHGs is a curve that is S-shaped and represents the

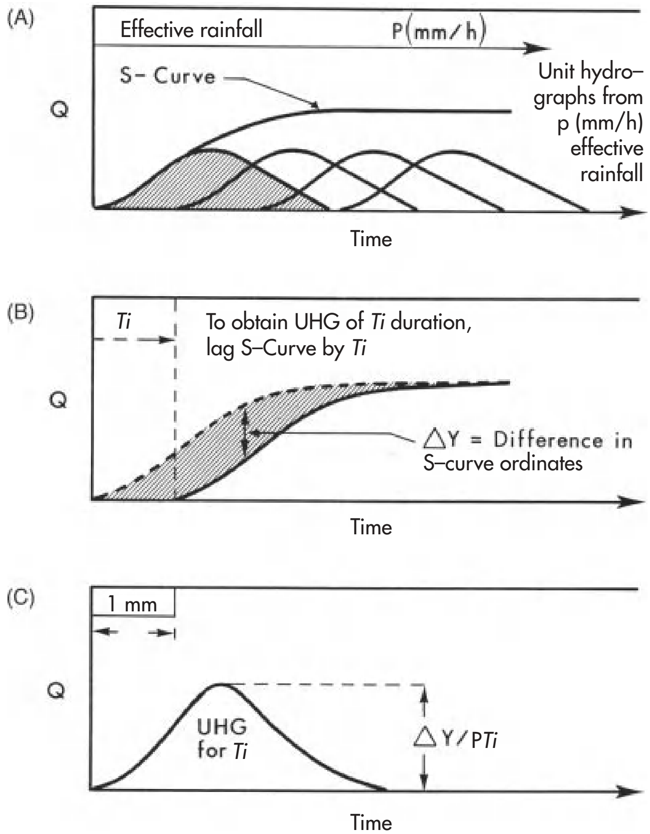

FIGURE 6.8. S-curve method of converting a UHG of one duration of *P* (mm/h) effective rainfall to another (T_i)

stormflow response expected for a rainstorm of infinite duration but at the same intensity of precipitation as the component UHGs. That is, the intensity would be 0.25 mm/h for a 4-hour UHG. Once the S-curve is constructed, a UHG of any duration can be obtained by lagging the S-curve by the desired UHG duration, subtracting the ordinates, and correcting for precipitation-intensity differences (see Fig. 6.8).

Before an empirical method such as a UHG analysis is applied, the method should be tested with observed data. The method can be assumed to be valid if historical streamflow responses can be reconstructed with a UHG model. One of the more difficult problems with the testing and applying of UHGs is determining the appropriate loss rates. Such losses are dependent largely on watershed characteristics and AMCs.

Loss Rate Analysis. Loss rates as used in the UHG method represent the rainfall or snowmelt that occurs but that does not contribute to stormflow. Early engineering hydrology textbooks often used the term's *loss rates* and *infiltration rates* interchangeably. This interchange of terms led to the idea that stormflow occurs only when infiltration capacities are exceeded and, therefore, results entirely from surface runoff. This situation is not necessarily the case on forested watersheds, however, where subsurface interflow often predominates. As applied in the UHG method, it is not necessary to equate loss rates with infiltration rates. Therefore, a more appropriate definition of "losses" would include precipitation stored and evaporated on vegetative surfaces (interception), precipitation stored in soils when soil-moisture deficits exist at the outset of a storm, or water that percolates to deep groundwater or is otherwise delayed so that it is not part of stormflow.

A loss rate function, therefore, need not approximate infiltration curves. The function can be approximated in most instances with either a constant loss rate (a Φ index) or a constantly diminishing loss rate. Importantly, loss rates have to be determined empirically for the entire watershed and generally relate to antecedent soil moisture conditions of a watershed.

Synthetic Unit Hydrographs. Application of the UHG method can be limited on many watersheds because both rainfall and streamflow data must be available as mentioned earlier. Because streamflow data are often not available, synthetic UHG methods have been developed. These methods consist of mathematical expressions that relate measurable watershed characteristics to UHG characteristics. Streamflow hydrographs for ungauged watersheds can be estimated with synthetic UHG models if loss rates can be approximated.

Numerous methods for developing synthetic UHGs are available, but they are not discussed in detail here. However, for general information, the Soil Conservation Service (SCS) Dimensionless UHG and several time–area-based approaches are presented in the *Unit Hydrograph Technical Manual of the National Weather Service*. The URL for this manual is http://www.nohrsc.nws.gov/technology/gis/uhg_manual.html (accessed July 21, 2011).

One example of a synthetic UHG is the SCS triangular hydrograph that couples the geometry of a triangle with time to peak estimates for a watershed (Fig. 6.9). The lag from

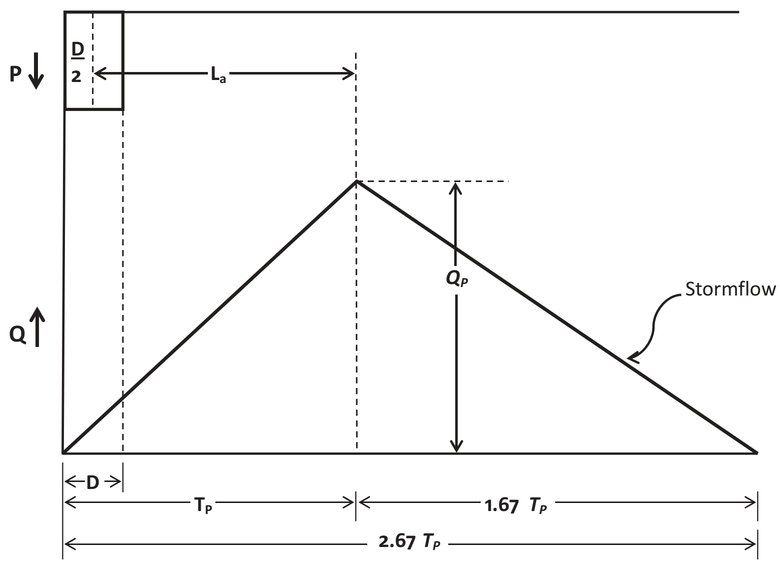

FIGURE 6.9. SCS triangular hydrograph (modified from Pilgrim and Cordery, 1993)

the centroid of rainfall excess (hyetograph) is assumed to be 0.6 T_c. The time to peak (T_p) is then determined to be 0.5D + 0.6 T_c, where D is the duration of rainfall excess. The volume of stormflow (Q) in the triangular hydrograph is determined as:

$$Q = 0.5Q_p \cdot T_b \tag{6.11}$$

where T_b is the base length of the hydrograph, assumed to be 2.67 T_p.

Peak discharge (Q_p) in ft^3/s can be approximated as:

$$Q_p = \frac{k_o A Q}{T_p} \tag{6.12}$$

where K_o is a constant (a commonly used value is 484 which means that 3/8 of the UHG volume is under the rising limb); A is the watershed area (mi^2); Q is the stormflow volume (in.); T_p is the time to peak (h); and T_c is the estimated time of concentration (h).

In applying Equation 6.12, K_o is not a universal constant. For steep mountainous terrain, K_o values could be close to 600, whereas in flat terrain such as in wetlands, K_o could be closer to 300. Determining values of K_o requires that synthetic hydrographs be compared to actual hydrographs when possible. The stormflow volume (Q) must also be known.

The Curve Number Method

The SCS-CN method (U.S. Soil Conservation Service, 1972) incorporates generalized loss rate and streamflow relationships developed in watershed studies throughout the United States. The following equation (in English units) is used to estimate the stormflow volume from a storm:

$$Q = \frac{(P - 0.2S_t)^2}{P + 0.8S_t} \tag{6.13}$$

where Q is stormflow (in.); P is rainfall (in.); S_t is watershed storage factor (in.); and $0.2S_t$ is equal to I_a, which is an initial loss that was a consensus value determined in the original development of the method.

Demonstration of the I_a 0.2 value for the initial loss is found only in the original document describing the CN method (U.S. Soil Conservation Service, 1972). However, after comparing rainfall-runoff data from 114 small mostly agricultural watersheds at 26 sites throughout the United States, Hawkins and Khojeini (2000) found the initial loss value more commonly near 0.05 rather than 0.2. The smaller initial loss value was apparent in situations of smaller rainfalls and lower CNs. Users of this method should consider these results.

A CN is related to S_t as follows:

$$CN = \frac{1000}{10 + S_t} \tag{6.14}$$

Soils, vegetation, and land-use characteristics are related to CNs that indicate the runoff potential for a specified rainfall event (Table 6.4). Soils are classified hydrologically

Table 6.4. **Runoff CNs for hydrologic soil-cover complexes for AMC II**

Cover			Hydrologic soil group			
Land use	**Treatment or practice**	**Hydrologic condition**	**A**	**B**	**C**	**D**
Fallow	Straight row		77	86	91	94
Row crops	Straight row	Poor	72	81	88	91
	Straight row	Good	67	78	85	89
	Contoured	Poor	70	79	84	88
	Contoured	Good	65	75	82	86
	Contoured and terraced	Poor	66	74	80	82
	Contoured and terraced	Good	62	71	78	81
Close-seeded legumes[a] or rotation meadow	Straight row	Poor	66	77	85	89
	Straight row	Good	58	72	81	85
	Contoured	Poor	64	75	83	85
	Contoured	Good	55	69	78	83
	Contoured and terraced	Poor	63	73	80	83
	Contoured and terraced	Good	51	67	76	80
Pasture or range		Poor	68	79	86	89
		Fair	49	69	79	84
		Good	39	61	74	80
	Contoured	Poor	47	67	81	88
	Contoured	Fair	25	59	75	83
	Contoured	Good	6	35	70	79
Meadow		Good	30	58	71	78
Woods		Poor	45	66	77	83
		Fair	36	60	73	79
		Good	25	55	70	77
Farmsteads			59	74	82	86
Roads (dirt)[b]			72	82	87	89
(hard surface)[b]			74	84	90	92

Source: U.S. Soil Conservation Service (1972).
[a] Close-drilled or broadcast.
[b] Including right-of-way.

into four groups

A: high infiltration rates; usually deep, well-drained sands and gravels with little silt or clay.
B: moderate infiltration rates; fine- to moderate-textured, well-structured soils such as light-sandy loams, silty loams.
C: below-average infiltration rates; moderate- to fine-textured, shallow soils; for example, clay loams.
D: very slow infiltration rates; usually clay soils or shallow soils with a hardpan near the surface.

Three AMCs are considered. Occurrence of these conditions is dependent on the amount of rainfall received five days before the storm of interest:

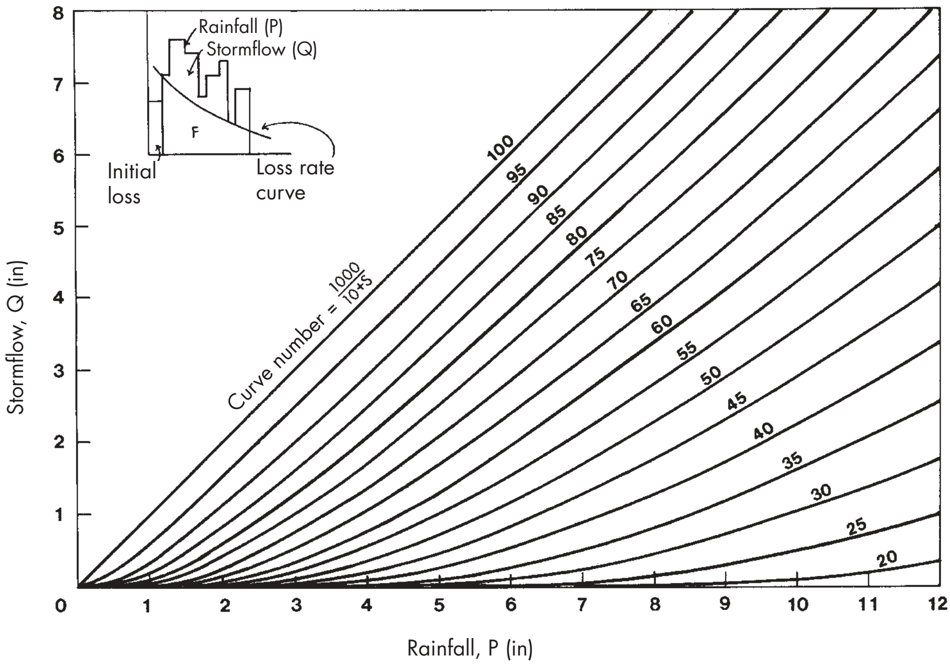

FIGURE 6.10. Rainfall-stormflow relationships for CNs (from U.S. Soil Conservation Service, 1972)

AMC I = dry, <0.6 in.
AMC II = near field capacity, 0.6 to 1.57 in.
AMC III = near saturation, >1.57 in.

Once the *CN* has been determined, rainfall is converted graphically to stormflow by the appropriate rainfall–stormflow relationship (Fig. 6.10). *CN* relationships should be determined for each hydrographically different region for the method to be valid. The relationships developed in the United States are generally applicable to small watersheds less than 13 km^2 in size with average slopes less than 30%. An application of the SCS method is outlined in Box 6.4.

Changes in stormflow associated with changes in the soil–vegetation complex are determined largely by the *CN* relationships (Hawkins and Ward, 1998). For example, changing from a good pasture condition within hydrologic soil group C with AMC II to a poor pasture condition changes the *CN* from 74 to 86. This change in turn would result in an increase in stormflow volume from 3.2 to 4.6 in. for a 6-in rain. In some instances, the time of concentration could also be altered, which would affect the peak and general shape of the stormflow hydrograph.

Analysis of Recession Flows

The *recession stage* of the hydrograph extends from the peak flow to the cessation of stormflow. It reflects the storage of water on a watershed, and, therefore, the effects of the vegetation, management, and land-use characteristics. Recession flows from different

Box 6.4

Application of the SCS Triangular Hydrograph and CN Method

A 4-in rain fell over a 4.6-mi^2 watershed in a good pasture condition, soils in group D, and an antecedent soil moisture condition II. The duration of rainfall was 6 hours. Time of concentration was estimated to be 5.5 hours. The stormflow and the corresponding peak flow are estimated as follows:

1. From Table 6.4, $CN = 80$.
2. As per Figure 6.10, for 4 in. of rain and CN 80, the stormflow is approximately 2.0 in.
3. Peak flow discharge for the watershed with a time of concentration (T_c) of 5.5 hours is determined as follows:

$$T_p = 0.5\,D + 0.6\,T_c$$
$$0.5\,(6) + 0.6\,(5.5) = 6.3 \text{ h}$$
$$Q_p = (484)\,(4.6)\,(2)/6.3 = 707 \text{ cfs}$$

watersheds can be characterized and compared by consolidating variables such as ET, soil moisture, and land-use patterns into one constant term for a recession-flow event.

Of the regression models available, the equation presented by Barnes (1939) is used widely in the United States. This equation is

$$Q_t = Q_0 k^t \tag{6.15}$$

where Q_t is the stormflow discharge after a time period t (cfs); Q_0 is the initial discharge (cfs); k is the recession constant per unit of time; and t is the time interval between Q_0 and Q_t in hours or days.

A linear form of this equation is obtained by a transformation onto semilogarithmic paper with discharge on the logarithmic scale and time on the arithmetic scale:

$$\log k = \frac{\log Q_t - \log Q_0}{t} \tag{6.16}$$

The slope k evaluated in Equation 6.16 is considered to be indicative of differences in recession flows. These changes are attributed collectively to the effects of vegetative cover, land-use patterns, season of the year, and soil moisture.

Streamflow data are frequently reported as daily averages. However, analysis of these averages can prevent the identification of individual hydrographs resulting from small storms not shown by the average streamflow for a day. Furthermore, the peaks identified by data points might not represent the actual peaks of the storms that caused the recession flows.

It is a frequent practice to arrange the recession flows chosen for analysis in a manner such that a composite recession flow is formed to represent the complete recession for a watershed (Singh, 1992). However, this procedure can be subject to judgmental errors and errors attributed to factors that contribute to the differences between short-event recessions. For example, *ET* losses vary greatly from time to time. Therefore, evaluation of the constant k for a recession based on a composite curve might not always be justified. Instead, the average recession flow might be evaluated with the variability measured as differences between slope constants of individual recession limbs.

Computer Simulation Models

Computer simulation models are representations of actual streamflow regimes and other hydrologic characteristics that allow people to study the functioning of watersheds, predict the response of streamflow regimes to varying climatic and management inputs, and obtain a better understanding of the general hydrologic impacts in doing so. The formulation, structure, and applications of computer simulation models are based largely on the *systems approach* to analysis. That is, an analysis of all of the interacting elements that function together for a purpose. Computer simulation models differ in terms of how and to what extent each component of the hydrologic process is included in their structures. A general discussion on the formulations, structures, and applications of computer simulation models of hydrologic systems including streamflow regimes is presented in Chapter 16.

Streamflow Routing

Stormflow hydrographs for a watershed become modified as the streamflow moves downstream through the channel. For example, the effect of a flood wave moving through a channel is an attenuation of the peak flow, which is the result of temporary water storage in the channel. Furthermore, the shape, speed, and magnitude of a flood wave change as it moves downstream and more drainage is accumulated. *Streamflow routing* is a mathematical procedure for predicting these changes. Streamflow routing is essential to predicting the changes in the magnitude of the peak flow and the corresponding stage of flow as a flood wave moves through the channel. A routing procedure is required if we attempt to determine how the changes in the streamflow hydrograph upstream become transferred to downstream locations. As streamflow enters larger order streams, the routing and combining of flows of all tributary streams need to be taken into account to determine the total hydrograph downstream.

Several methods are available for routing a streamflow through a channel. Hydraulic routing takes into account the channel features and computes flow along several cross sections of a channel over time. Hydrologic routing is an empirical method of routing flow over time though a channel using a generalized relationship for the entire channel reach. Whereas it is not a purpose of this book to discuss the array of routing procedures used in hydrology, a generalized hydrologic routing method is described for illustrative purposes. The generalized hydrologic routing method described is referred to as a reservoir-routing method or lumped-flow-routing procedure as described by Fread (1993). This method is based on the conservation of mass principle that states:

$$I(t) - O(t) = ds/\mathrm{d}t \qquad \textbf{(6.17)}$$

where $I(t)$ is the inflow at the channel reach as a function of time; $O(t)$ is the outflow from the channel reach as a function of time; and ds/dt is the time rate of change of storage in the channel reach.

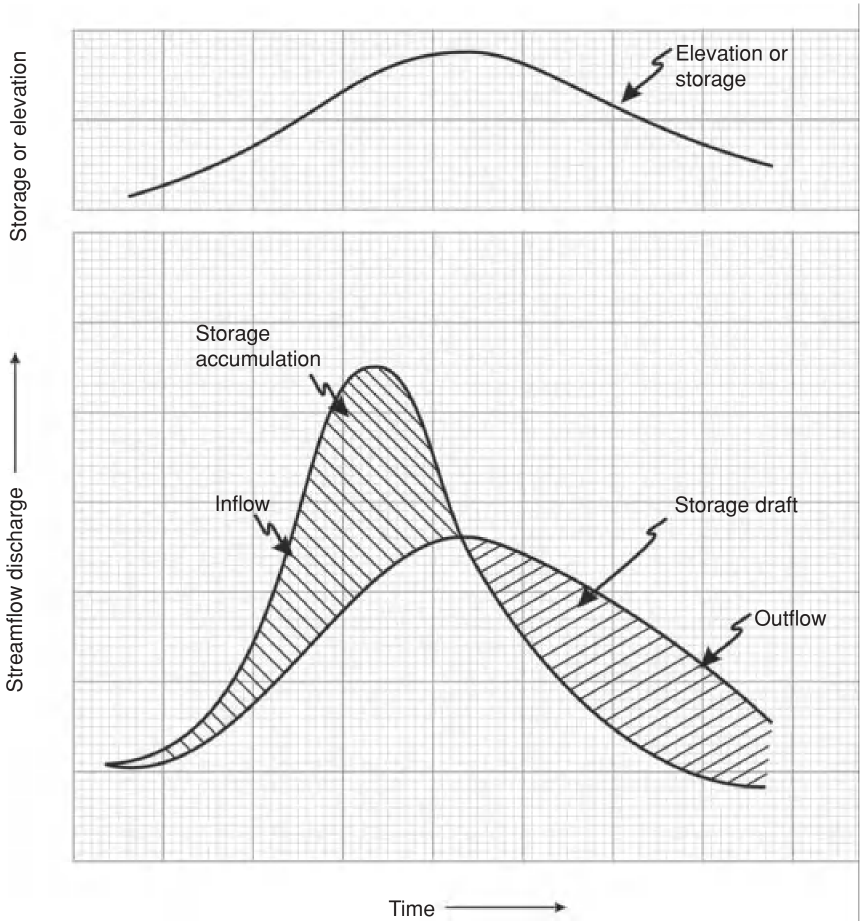

FIGURE 6.11. Example of reservoir-routing method for streamflow routing

The storage s is a function of both inflow and outflow and can be related to either or both by an empirical storage function. If it is assumed that the water surface is level throughout the reach (as in a lake or reservoir), storage can be empirically related to outflow or the surface elevation (h), that is, $s = \mathrm{f}(h)$. A schematic illustration of this routing procedure is shown in Figure 6.11.

The generalized hydrologic-routing method described above is the basis for several mathematical-routing methods. Discussions of these routing applications can be found in Singh (1992), Fread (1993), Brutsaert (2005), and others.

Streamflow-Frequency Analysis

A frequency analysis of streamflow events is undertaken for purposes such as determining the chance or probability of extreme high- or low-streamflow events. For example, streamflow-frequency analysis is used to delineate a floodplain according to the flood risk and to determine conservation storage requirements of a reservoir. Streamflow *frequency* *curves* as the end product of such analysis are an expression of the occurrence of a streamflow characteristic such as a peak flow of a specified magnitude on a probability basis. The frequency with which the characteristic is equaled or exceeded is determined from a frequency curve.

The purpose of a frequency analysis should be specified initially in developing a frequency curve. A frequency analysis might be developed to determine storage needs for low-flow augmentation or to estimate the chance of experiencing a critical low-flow value for a particular time duration. The specified purpose will determine the specific type of data that are required. Important considerations include:

1. The appropriate streamflow characteristic needs to be identified, for example, instantaneous peak discharges of interest (floodplain delineation), daily flood flow volumes of interest (flood storage analysis), or minimum daily flows to determine reservoir storage required to meet low-flow augmentation releases.
2. The data used in the frequency analysis must represent a measure of the same aspect of each hydrologic event. That is, mean daily peak discharges cannot be analyzed together with instantaneous peak discharges.
3. Streamflow data being analyzed should be "controlled" by a uniform set of hydrologic and operational factors. Natural streamflows cannot be analyzed with flows that have been modified by reservoir operations. Likewise, peak discharges from snowmelt cannot be mixed with peak discharges from rainfall.
4. A *partial-duration series* analysis can be used to obtain a better definition of the frequency curve than an *annual series* if only a few years of streamflow records are available (see Fig. 6.12). However, this analysis does not necessarily make the frequency curve more reliable.

The annual-series approach uses only the extreme annual event such as the largest peak flow for a peak-discharge analysis from each year's streamflow record, whereas the partial-duration series uses all independent events above a specified base level in the analysis. For a peak-discharge analysis, therefore, an annual-series analysis ignores the second-highest

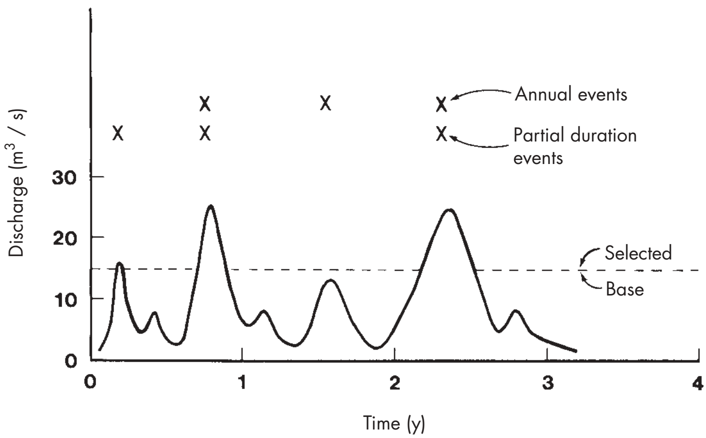

FIGURE 6.12. Annual- and partial-duration series approaches to peak flood analysis

peak discharge in a year (illustrated by 16 m^3/s in year 1 in Figure 6.12), which can be higher than the highest peak discharge in another year (14 m^3/s in year 2 in Figure 6.12). Including all major events in a partial-duration series can better define a frequency curve when only a few years of data are available. As the number of years of record increases, however, the frequency curve developed by either method will show essentially the same frequency relationship for the region of the curve that defines events of low probability. However, frequency curves developed by these two methods are interpreted differently. For an annual series, the *exceedance frequency* for a specified magnitude, that is, the frequency by which a value is equaled or exceeded, is the number of *years* that magnitude will be exceeded per 100 years. The interpretation for a partial-duration series is the number of events that will exceed a certain magnitude per 100 years.

Either a graphical or an analytical analysis can be used to develop frequency curves once the necessary data sets are available.

Graphical-Frequency Analysis

A graphical-frequency analysis involves calculating the cumulative probabilities for ranked events and then drawing the frequency curve through the data points. Several methods are available to calculate cumulative probabilities. The median formula (Beard, 1962) is commonly used for peak-discharge frequency curves:

$$P = \frac{m - 0.3}{N + 0.4} \tag{6.18}$$

where P is the cumulative probability (plotting position) for the event ranked m in N number of years of record.

The cumulative probability for the least severe events ($p < 0.5$) can be calculated as:

$$P = \frac{2m - 1}{2N} \tag{6.19}$$

The graphical-frequency analysis does not assume a statistical distribution but is quick and easy to apply. Since the frequency curve is hand drawn, one portion of a curve can be weighted more than another. It is usually recommended that the cumulative probabilities (plotting positions) determined by the graphical method be plotted, even if the analytical method described below is used to define the curve.

Analytical-Frequency Analysis

An analytical-frequency analysis requires that sample data follow some theoretical-frequency distribution. For example, The Interagency Advisory Committee on Water Data recommends the *log Pearson Type III distribution* for the analytical development of peak-discharge frequency curves. Guidelines for applying this method follow Flynn et al. (2006) as per the website: http://pubs.usgs.gov/tm/2006/tm4b4/ (accessed November 29, 2011).

The method entails the following steps:

1. The data are transformed by taking the logarithms of peak discharges.
2. The mean peak discharge (first moment) is calculated; this corresponds to the 50% probability of exceedance.

3. The standard deviation (second moment) is calculated; this represents the slope of the frequency curve plotted on log-probability graph paper.
4. The skew coefficient (third moment), an index of non-normality, is then calculated; this represents the curvature in the frequency curve.
5. Adjustments are then made for small numbers of events. The skew coefficient is usually unreliable for short records; a regionally derived skew coefficient is recommended for fewer than 25 years of records. A weighted skew should be used if 25–100 years of records are available. The skew derived from one station should be used only if more than 100 years of records are available.
6. The frequency curve for the observed annual peaks (Q) is then determined for selected exceedance probabilities (P) by the equation:

$$Log\ Q = \bar{x} + k_s s \tag{6.20}$$

where $\bar{x}$ is the mean of log Q (m^3/s); k_s is the factor that is a function of the skew coefficient and a selected exceedance probability; s is the standard deviation of log Q (m^3/s).
7. Confidence limits can be calculated and plotted for the frequency curve as a final step.

The analytical-frequency analysis has several advantages over the graphical method including

1. The same curve is always calculated from the same data, which makes the analysis more objective and consistent.
2. The reliability of the curve can be estimated by calculating confidence intervals.
3. The calculation of statistics by regional analysis described later in this chapter allows for the development of frequency curves at ungauged sites.

An example of a plotted-analytical-frequency curve with the corresponding graphical-plotting points is presented in Figure 6.13.

Frequency Analysis on Ungauged Watersheds

Frequency curves for ungauged watershed can be approximated by performing a regional analysis or modeling the streamflow response using more readily available and longer-record precipitation data. Either of these two approaches can also be used to extend the streamflow record for locations with only a few years of data.

A regional analysis is a statistical approach in which generalized equations, graphical relationships, or maps are developed to estimate hydrologic information at ungauged sites (USACE, 1975). Streamflow-frequency characteristics can be estimated for ungauged watersheds that are within the same climatic region as the gauged watersheds employed. Any pertinent information within the region should be used to relate watershed characteristics to hydrologic characteristics. For example, a regional analysis can be used to estimate runoff coefficients or peak discharge in cubic meters per second per square kilometer associated with a specified recurrence interval or to estimate the constants needed to execute a complex hydrologic model.

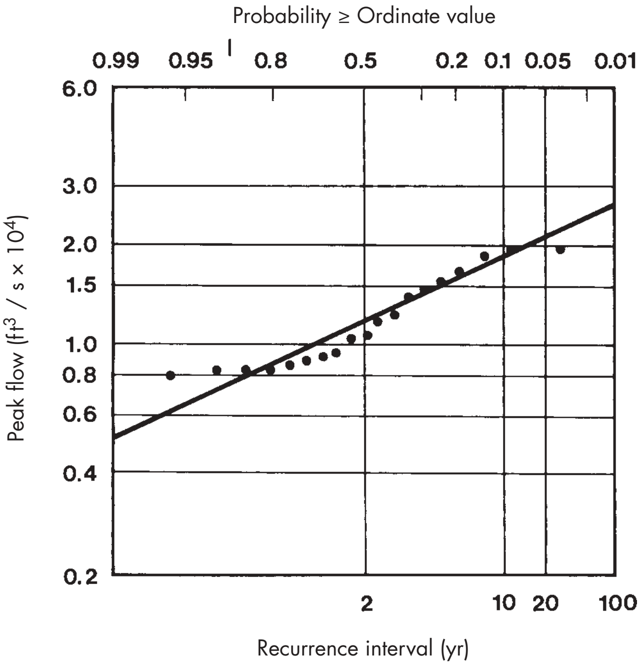

FIGURE 6.13. Analytical-frequency curve with graphical plotting positions of annual peak discharge for the Little North Santiam River near Mehama, Oregon (1932–1950)

Equations and maps are developed that allow the derivation of exceedance-frequency curves for ungauged areas in the following manner:

1. Select components of interest such as the mean annual peakflow, 100-year recurrence-interval peakflow, etc.
2. Select explanatory variables (characteristics) of gauged watersheds such as drainage area, watershed slope, and percentage of area covered by lakes or wetlands.
3. Derive prediction equations with single- or multiple-linear regression analyses.
4. Map and explain the residual errors that are the differences between calculated and observed values at gauged sites.
5. Determine frequency characteristics for ungauged locations by applying the regression equation with adjustments as indicated by the mapped residual errors.

The residual errors constitute unexplained variance in the statistical analysis. Mapping the residuals can sometimes indicate a relationship between the magnitude of the residual error and watershed variables that were not included in the regression analysis such as vegetative type, soils, or land use. The results of the analyses should be considered a rough estimate of the true hydrologic characteristics in any case.

The modeling approach for developing streamflow records for frequency analysis involves the following steps:

1. Calibrate the hydrologic simulation model using gauged data from similar watersheds to establish the constants or coefficients used by the model. Hydrologic judgment and

knowledge of the model structure and the regional hydrology are essential to this process.
2. Adapt the model to the ungauged watersheds using measurable data and estimated parameters.
3. Enter precipitation data from the nearest long-term gauge or generate a long-term record using a stochastic model for the region. A standard frequency analysis can then be conducted on the resulting streamflow output from the model.

Comments on Frequency Analysis

Several issues arise when performing streamflow-frequency analysis. Streamflow data are lacking for most small streams and streams in remote areas. Therefore, developing a regional analysis can be a challenge. When these data are available, the record periods are often short. The uncertainty associated with frequency curves for these watersheds can be substantial.

A level of uncertainty can also exist with relatively long-term streamflow records because even these records are only a sample of the possible events that have occurred in the past or might occur in the future. Streamflow patterns vary with climatic variability and change. Frequency curves developed from the past 50 years might not represent the next 50 years. The tendency in these situations is to guess high for flood frequencies to provide conservative answers. In all cases, the consequences of substantial errors in a calculated frequency curve should be the guide for determining the acceptable risk.

Watersheds are also undergoing changes that are likely to affect the streamflow responses to these changes to some degree. Timber harvesting, livestock grazing, occurrence of wildfire, road construction, and urbanization can all affect streamflow response of watersheds (see Chapter 12). Such changes are sometimes abrupt, and, therefore, the corresponding change in the hydrologic regime is often identified readily. More subtle long-term changes that are cumulative are perhaps the most commonly encountered and also the most troublesome in trying to relate watershed changes to changes in streamflow-frequency curves.

Whenever watershed changes are such that a change in streamflow response is expected, extreme care must be exercised in a streamflow-frequency analysis. One possible solution to this situation is to use a computer simulation model with sufficient sensitivity to predict both the land-use change and associated streamflow response. The frequency analysis can then be performed with the simulated data as discussed above.

Summary and Learning Points

The streamflow response of a watershed as a result of rainfall or snowmelt-runoff events is the integrated effect of many factors. Streamflow data are necessary for many operational purposes including streamflow forecasting, determining available water supplies, and developing streamflow-frequency relationships for planners and managers. Hydrologists and watershed managers also need tools to predict runoff and streamflow discharge for a variety of purposes such as designing of water storage and conveyance systems or predicting how changing precipitation patterns and land-use activities will alter runoff and streamflow response. By now, therefore, you should be able to:

1. Determine streamflow discharge from knowledge of streamflow velocity and cross-sectional area of the stream.
2. Describe different ways in which a streamflow can be measured and discuss the advantages and disadvantages of each.
3. Define the respective terms in the Manning and Chezy equations and present an example of how these equations can be used to estimate peak discharge of a streamflow event that was not measured directly.
4. Define a UHG and explain the conditions under which it can be applied to estimate stormflow.
5. Select a method to accomplish objectives such as:
   - determine the peak discharge and stormflow volume for a watershed that has streamflow records;
   - examine the effects of different land uses on stormflow peak and volume; and
   - determine the probability that a peak discharge of a certain magnitude will be equaled or exceeded in a specified number of years.

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CHAPTER 7 *Groundwater and Groundwater–Surface Water Exchange*

INTRODUCTION

Water in saturated zones lying beneath the soil surface is groundwater. Groundwater comprises more than 97% of all of the liquid freshwater on earth. More than one-half of the world population depends on groundwater. Groundwater contributes about 30% of all streamflow in the USA. Furthermore, about one-third of public water supplies in the USA comes from groundwater (http://ga.water.usgs.gov/edu/wugw.htmlee). Although groundwater is an important source of freshwater, it does not always occur where it is most needed and is sometimes difficult to extract. Without appropriate management, large quantities of this valuable resource can become unusable because of contamination or by mining, which is pumping to the point that further extraction is not feasible economically.

Groundwater is an essential input to many streams, lakes, and wetlands (Winter et al., 1998). Groundwater is essential for permanent land-locked lakes and perennial flows of water. Isolated lakes can derive nearly 50% of their volume from groundwater in some regions (Winter, 1997). These lakes typically have an up-gradient side of the lake where groundwater flows into the lake and a down-gradient portion of the lake water recharges back into the groundwater (Magner and Regan, 1994; Magner et al., 2001). Wetlands are primarily surface expressions of a water table or an upwelling of groundwater in discharge zones.

The purpose of this chapter is to familiarize the reader with the characteristics and properties of groundwater in a general context and focus attention on the linkages between groundwater and streams, lakes, and wetlands.

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

GROUNDWATER

The basic concepts, the storage and movement, and the development of groundwater resources and the effects of vegetation on shallow groundwater aquifers are discussed in this section.

Basic Concepts

Some perceive groundwater to occur as vast underground lakes or rivers, but for the most part groundwater occurs in voids between soil and rock particles in a *zone of saturation*. To understand how the zone of saturation is formed, we must consider the forces that govern the downward movement of water in the soil as discussed in Chapter 5. This downward movement of water past the root zone continues through pores and fractures in the soil, parent material, and underlying rock. In some geologic settings, pores become fewer and smaller with increasing depth depending on geologic origin, generating more resistance to flow. In other geologic settings such as karst limestone, water flows through fractures and caves at varying depths below the ground surface (White, 2007). In either case, the water is no longer subject to evaporation or transpiration losses as it moves into deeper zones. Although impervious or restrictive layers (or strata) can exist at various depths below the soil surface, porosity generally diminishes at depths below 600 m and, therefore, wells that are drilled to such depths find little water (Baldwin and McGuinnes, 1963). However, the depth where water movement from above is slowed and allowed to accumulate delineates the *phreatic zone* or the zone of saturation.

The *zone of aeration* is that part of the profile that occurs between the soil surface and the top of the zone of saturation (Fig. 7.1). The zone of aeration consists of the soil-water zone, which extends through the rooting zone, and the *vadose zone* that extends from the soil-water zone to the capillary fringe. The top of the zone of saturation where the water potential is zero is the *water table*. It is measured by the elevation of water surfaces in wells that penetrate into the zone of saturation. Immediately above the water table is the *capillary fringe*, a zone in which water from the zone of saturation is pulled up by capillary forces into the zone of aeration. This capillary fringe has a negative water potential and its irregular position varies with changes in water table elevation. The height of the capillary fringe above the water table is determined by the type of sediment matrix. It is insignificant in coarse-grained sediments but can be several centimeters high in silts and clays.

The above terminology should not suggest that the zone of aeration cannot become saturated. It can frequently become saturated in times of soil-water excess. A distinguishing feature of the zone of aeration is that the saturated conditions are variable. It should also be emphasized that the water table is not a static surface. The elevation of a water table moves up and down in response to changing precipitation and evapotranspiration (ET) patterns (see Fig. 7.1). During a wet season, springs occur where the water table comes in contact with the soil surface and the groundwater system can discharge water into streams, lakes, and wetlands.

A break in land-surface slope often results in groundwater *resurgence* or *discharge* into surface water. If the surface water is a lake or wetland, however, it is difficult to observe the groundwater resurgence unless iron-bacteria or an organic sheen is present on the water surface. An indicator of groundwater resurgence is a lake or wetland outlet where the flow of

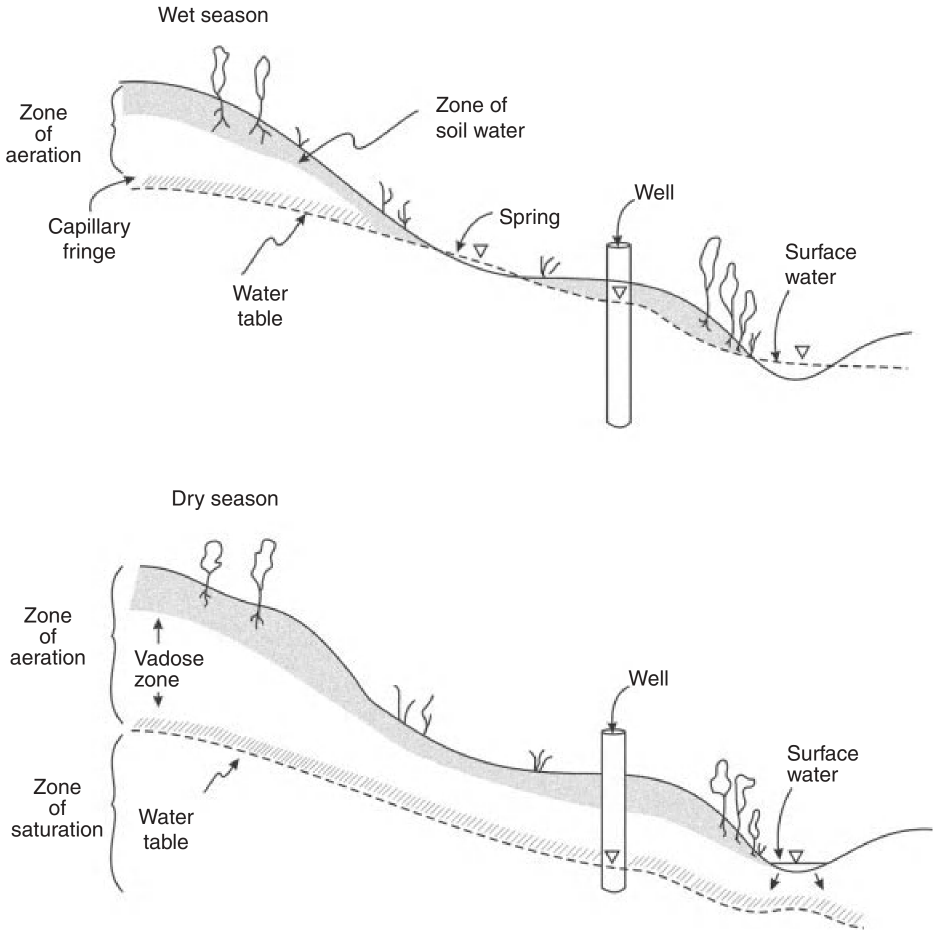

FIGURE 7.1. Groundwater characteristics and water table changes from a wet to a dry season

water can be measured. These areas are known as *headwater* or *source water areas* because they represent the areas of subsurface water convergence. Convergence of subsurface water that results in the flows of surface water is considered a headwater because surface water can begin to flow in a stream channel from these areas that are sometimes called *variable source areas* (Dunne and Leopold, 1978), as discussed in Chapter 5.

Streams that are groundwater fed are called *effluent streams* and usually are perennial. In dry periods, the water table can drop below the streambed to create situations where springs no longer flow and where streams are no longer fed by groundwater. Streams that lose flow to the groundwater are *influent streams* (see Fig. 7.1 and Fig. 5.9c). In some settings, *losing-reach* streams function more like a semi-impervious canal, routing water from a lake or wetland or geologic strata to a lower elevation with some seepage loss along the path.

Storage and Movement of Groundwater

Groundwater is found in many different types of soil and rock strata. It can occur between individual soil or rock particles, in rock fracture openings, and in solution openings formed when water dissolves mineral constituents in the rock strata, leaving a void. The amount of groundwater stored and released from the water-bearing strata depends on the porosity, the size of pore spaces, and the continuity of pores. Water-bearing porous soil or rock strata that yield significant amounts of water to wells are called *aquifers*. Soil or rock strata of shale, slate, or highly compacted and cohesive soil such as basal till are barriers to downward water movement and, therefore, are effectively impermeable or semi-impermeable and generally referred to as *aquicludes*. Slowly permeable geologic strata that retard the movement of groundwater such as lacustrine clays, fractured mudstone, and certain sandstones are *aquitards*. These terms are broadly used to generally describe regional geologic strata because natural systems tend to be complex. Geologic complexity that occurs over varying spatial scales is referred to as *heterogeneity*.

An aquifer can be an underground lens of sand or gravel, a layer of sandstone, a region of highly fractured rock including granite, or a layer of cavernous limestone. An aquifer can be from a meter to hundreds of meters thick and can underlie a few hectares or thousands of square kilometers. For example, the Ogallala aquifer underlies about 450,000 km^2 in the Midwestern region of the USA (http://www.waterencyclopedia.com/Oc-Po/Ogallala-quifer.html).

Porosity, the total void space between the grains and in the cracks of aquifers and solution cavities that can fill with water, is defined in terms of the percentage of pore space as

$$\text{Porosity} = \frac{100V_\text{v}}{V_\text{t}} \tag{7.1}$$

where V_v is the volume of void space in a unit volume of rock or soil; and V_t is the total volume of earth material including void space.

The effective porosity is the ratio of the void space through which water can flow to the total volume. If all the grains in a consolidated or unconsolidated material are about the same size and well sorted, the spaces between the grains account for a large part of the total volume. If the grains are sorted poorly, however, the larger pores can fill with smaller particles rather than water. Well-sorted materials, therefore, tend to hold more water than materials that are sorted poorly. Porosity ranges from 10% to 20% for glacial till that is poorly sorted material with a range of particle sizes derived by glacial activity, from 25% to 50% for well-sorted sands or gravels, and from 33% to 60% for clay.

The pores must be connected to each other if water is to move through a soil or rock stratum. If the pores are interconnected and sufficient in size to allow water to move freely, the soil or rock is *permeable*. Aquifers that contain small pores or weakly connected pores yield only small amounts of water even if their total porosity is high.

Groundwater flows through a soil or rock strata in response to hydraulic-head gradients and, in doing so, follow the pathway of least resistance. These flows will move through permeable materials and around impermeable ones. The complexity of the geology will influence how water moves, that is, increased amounts of folding, uplifting, and fracturing of strata will produce complex pathways of water movement. Fracture flow can occur in *aquicludes* and *aquitards* because of near-surface geologic processes that shift the earth's crust such as tectonics or glacial rebound.

Fractures are formed when a resistant material breaks or shatters under stress. These subsurface features can be discerned over large scales by examining aerial photos that reveal linear surface indicators known as lineaments. In South Africa, Dolerite dikes can be seen on aerial photos, guiding hydrogeologists in water development decisions. Water and entrained pollutants can move rapidly through the subsurface when fracture flow occurs (Field, 1992). This rapid movement of water occurs because the water bypasses primary pores and moves through a series of interconnected open spaces that characterize a situation known as *secondary porosity*. Except in karst terrain, secondary porosity can become limited, resulting in slower movement of groundwater and increased hydraulic residence time in a given strata.

As groundwater moves from high-elevation recharge zones toward river basins, the hydraulic residence time or time in contact with the aquifer minerals changes the chemical composition of the water. Depending on the streamline pathway of flow, groundwater can flow slowly for hundreds of kilometers before emerging as a natural spring, seeping into a stream, being tapped by a well, or emerging into an ocean. Unless extracted by a well, however, some groundwater is *connate*; that is, it is out of circulation within the hydrologic cycle (see Box 7.1).

Unconfined and Confined Aquifers

Aquifers that contain water that is in direct contact with the atmosphere through porous material are called *unconfined aquifers*. The groundwater system illustrated in Figure 7.1 is unconfined with the soil system immediately above the water table allowing for the exchange of gases and water. In contrast, a *confined aquifer* is separated from the atmosphere by an impermeable layer, *aquiclude* or *aquitard* (Fig. 7.2). A confining stratum often forms a *perched* water table; perched meaning above the semi-regional or regional flow of groundwater. An unconfined aquifer can become a confined aquifer at some distance from the recharge area (Fig. 7.2).

Confined aquifers, also called *artesian aquifers*, contain water under pressure that in some cases is sufficient to produce freely flowing wells. Water pressure (P), or pressure potential, is a function of the height of the water column at a point (h_p), the density of water (p), and the force of gravity (g). For a system without energy loss due to flow friction, the water pressure can be approximated by

$$P = \rho\, g\, h_p \tag{7.2}$$

Pressure is directly proportional to the height of the water column above some point in the system. The total hydraulic head (h_t) includes the water pressure from that point down to an arbitrary but stable reference datum:

$$h_t = z + h_p \tag{7.3}$$

These components are illustrated in Figure 7.3. The difference in total hydraulic head from one point to another creates the hydraulic gradient dh_t/dx where x is the distance between points. In unconfined aquifers, the elevation of the water surface measured by wells can be used to construct a water table contour map, which is similar to a surface contour map for surface water elevation. The direction of groundwater flow, or *streamlines of flow*, can be

Box 7.1

Connate Water in the Red River Valley of the North, USA (Magner et al., 2001)

In the Red River of the North valley, there is an increasing demand for agricultural, industrial, and municipal water use. This has led to concern about water availability, water-level decline, and degraded water quality in the Buffalo River watershed and the Buffalo aquifer. Carbon-14 ($\delta^{14}C$), tritium ($\delta^{3}H$), oxygen-l8 ($\delta^{18}O$), and deuterium (δD) were sampled from selected surface waters and domestic and municipal wells of varying depth. Nested well clusters and the Buffalo River were sampled in 1993 along an east–west watershed transact. Intensive isotopic sampling occurred in August 1993 during a pump test of Moorhead Municipal Well One. Results showed that groundwater in the Buffalo River watershed entered the subsurface prior to 1953 except for shallow moraine groundwater. Groundwater below the lake-plain was at least 8000 years old and was likely of glacial origin based on light $\delta^{18}O$ values below the lake-plain that supported the $\delta^{14}C$ results. Groundwater $\delta^{18}O$ values from the moraine were heavier than deep lake-plain groundwater, yet lighter than the average annual precipitation for the region, suggesting preferential snowmelt recharge compared to summer rainfall sources. The Buffalo aquifer was the sole source of water to Moorhead Municipal Well One during a 10-day pump test. Tritium concentrations in water from the Buffalo River, Buffalo aquifer, and surrounding lake-plain aquifers indicated that recharge to the Buffalo aquifer occurred both prior to and after 1953, the time of peak tritium concentrations in the atmosphere. These data suggest that though the Buffalo aquifer has a long hydraulic residence time, it can be susceptible to contamination in the future from present-day land-use practices.

determined by constructing lines perpendicular to the water table contours from higher to lower elevation contours.

The piezometric or *potentiometric* surface of an artesian aquifer indicates the imaginary level of hydraulic head to which groundwater will rise in wells drilled into the confined aquifer (see Fig. 7.2). The potentiometric surface declines because of friction losses between points. But when the land surface falls below the potentiometric surface, water will flow from the well without pumping, that is, an artesian or flowing well. Therefore, artesian pressure is the result of the actual water table in the downstream discharge area being at a much lower level than in the upstream recharge area and its suppression by confining layers above the aquifer.

As with unconfined aquifers, points of equal potentiometric head can be connected forming contours that are used to construct a *potentiometric*-surface map. Such a map

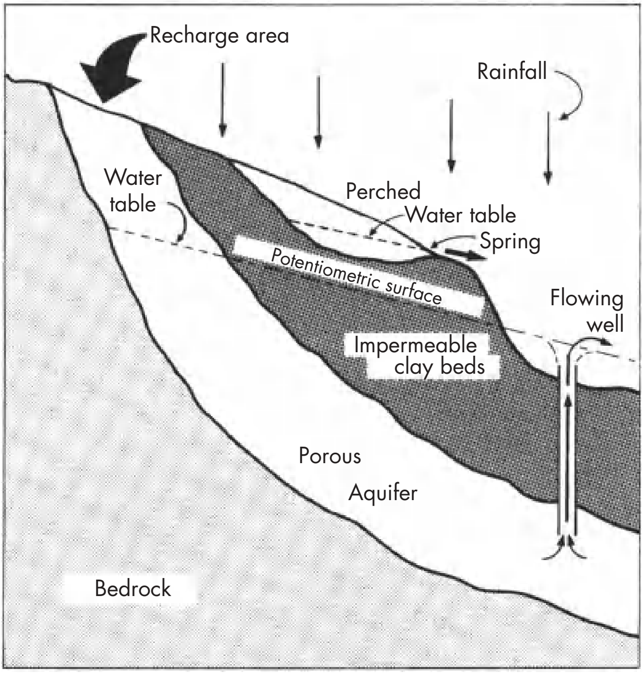

FIGURE 7.2. Artesian (confined) aquifer and recharge area with a perched water table above an impermeable layer (adapted from Baldwin and McGuinness, 1963)

represents the slopes of the potentiometric surface and indicates the direction of groundwater flow in artesian aquifers.

Aquifer Characteristics

When considering the development of groundwater for pumping, certain characteristics of the aquifer(s) from which the groundwater is to be extracted need to be understood. An important characteristic is the *transmissivity* of an aquifer, which is the amount of water

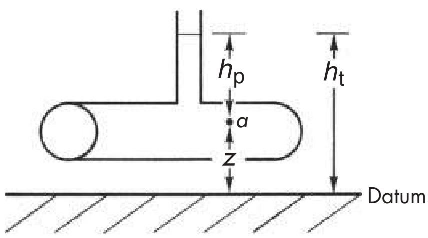

FIGURE 7.3. Pressure head (h_p), elevation head (z), and total hydraulic head (h_t) of water. Elevation head (z) is the distance from an arbitrary but stable reference datum to a point (a) where the pressure head (h_p) is measured

that can flow horizontally through the saturated thickness of the aquifer under a hydraulic gradient of 1 m/m. Transmissivity is defined as

$$T_r = bk_v \tag{7.4}$$

where T_r is the transmissivity (m^2/unit time); b is the saturated thickness (m); and k_v is the hydraulic conductivity of the aquifer (m/unit time).

Any time the hydraulic head in a saturated aquifer changes, groundwater will be either stored or discharged. *Storativity* is the volume of water that is either stored or discharged from a saturated aquifer per unit surface area per unit change in head. The storativity characteristic of an aquifer is related to the specific yield of the soil or rock material that constitutes the aquifer. *Specific yield* (S_y) is the ratio of the volume of water that can drain freely from saturated earth material due to the force of gravity to the total volume of the earth material.

The amount of water discharged from an aquifer can be determined from Darcy's law (Chapter 5).

An application of Darcy's law (Equation 7.5) to determine the discharge of water through a well-sorted aquifer of gravel is shown in Box 7.2. The discharge of flow from the aquifer is dependent on the cross-sectional area through which flow occurs (A), the hydraulic conductivity of the material constituting the aquifer (k_v), and the hydraulic gradient (dh_t/dx). The value of k_v is dependent on the properties of the porous medium and the fluid passing through it – the more viscous the fluid, the lower the k_v. Examples of hydraulic conductivities of the earth material for pure water at a temperature of 15.6°C are listed in Table 7.1.

$$Q = k_v A \frac{dh_t}{dx} \tag{7.5}$$

Box 7.2

Application of Darcy's Law

The problem presented is to determine the discharge of flow through a well-sorted gravel aquifer, given that $k = 0.01$ cm/sec, the change in head is 1 m over a distance of 1000 m, and the cross-sectional area of the aquifer is 500 m^2.

$$\begin{aligned} Q &= k_v\, A\, [dh_t/dx] \\ &= (0.01\ \text{cm/sec})\,(500\ \text{m}^2)\left(\frac{10{,}000\ \text{cm}^2}{1\ \text{m}^2}\right)(0.001\ \text{m/m}) \\ &= 50\ \text{cm}^3/\text{sec} \\ &= 4.32\ \text{m}^3/\text{day} \end{aligned}$$

TABLE 7.1. Examples of hydraulic conductivities for unconsolidated sediments (pure water, 15.6°C)

Material	Hydraulic conductivity (cm/sec)
Well-sorted gravel	10^{-2} to 1
Well-sorted sands, glacial outwash	10^{-3} to 10^{-2}
Silty sands, fine sands	10^{-5} to 10^{-3}
Silt, sandy silts	10^{-6} to 10^{-4}
Clay	10^{-9} to 10^{-6}

Source: Adapted from Fetter (2001).

Groundwater Development

Groundwater can provide sources of freshwater to meet the demands for municipal-industrial uses, irrigation, and other uses on watersheds. Assessing the potential for groundwater development requires knowledge of the local geology and aquifers. While surface features offer limited insight into the location, depth, and extent of water-bearing material or strata, geologic maps can be used to help a groundwater hydrologist in identifying potentially productive water-bearing strata by

- examining the direction and degree of dipping strata;
- locating faults and fracture zones; and
- determining the stratigraphy of rocks with water-bearing and hydraulic characteristics.

Information obtained from geologic maps can also be used to determine whether special extraction techniques such as horizontal wells are appropriate. For example, areas that have old lava flows often exhibit considerable vertical development of secondary openings such as lava tubes and fissures caused by escaping gases. Horizontal wells increase the chances of intercepting these larger water-bearing pores that are usually not widespread and are difficult to locate by vertical borehole drilling.

Opportunities for groundwater development generally increase when moving from upland watershed areas to lower basins and floodplains. Extensive and high-yielding aquifers occur in most major river valleys and alluvial plains. On a smaller scale, the same features can be indicative or important sources of groundwater in upland areas. Small valleys in upland areas and areas associated with stream-channel systems often contain locally high water-yielding deposits of alluvium. Although usually not extensive, these deposits can provide water during dry periods or be a backup for other water supply systems.

With sufficient aquifers and proper well location, groundwater can supply water for people, their livestock, and irrigation of agricultural lands. The amount of water that can be supplied from wells depends on

- the types of rocks underlying the area and the degree of weathering;
- the presence of faults or fracture zones; and
- the extent of unconsolidated sands and gravels occurring as alluvium or located below stream channels.

TABLE 7.2. Water-bearing and yield characteristics of some common aquifers

		Well yields	
Type of material	**Specific yield (%)**	**GPM**	**L/sec**
Metamorphic/plutonic igneous	0–25	10–25	0.6–1.6
Volcanic	Variable	<1500	<95
Granite	<1	Neg.	Neg.
Sedimentary rocks			
Shales/claystones	0–5	<5	<0.3
Sandstones	8	5–250	0.3–16
Limestone, solid	2	Neg. (–5)	Neg. (–0.3)
Limestone with solution cavities	Variable	>2000	>126
Unconsolidated deposits			
Clay	0–5	Neg.	Neg.
Sand/gravel	10–35	10 to >3000	0.6 to >189

Source: Adapted from Davis and DeWiest (1966) and Fetter (2001).
GPM, gallons/min; Neg., negligible.

However, the yield of wells constructed in consolidated and unconsolidated materials vary considerably as shown in Table 7.2.

Wells

There are many types of wells, including those that are hand dug, driven into an aquifer using well points, or drilled with a cable tool or a rotosonic drill rig. The specialized drill head of a rotosonic drill imparts high-frequency vibrations into a steel drill pipe, creating a drilling action that allows the retrieval of continuous, undisturbed cores.

Dug wells are not necessarily deep but are often lined to keep the sides from falling in. Pipes are pushed into shallow gravel and sand aquifers usually less than 20 m below the surface in developing driven wells. Such wells are simple and cheap to install. Drilled wells differ from a dug well in that the hole is made with a drilling rig that enables deeper wells to be constructed as described in Box 7.3.

To test the performance or yield of a well, the water level in the well is measured first and then pumped at a steady rate. The water level will drop quickly at first and then more slowly as the rate at which water is flowing into the well approaches the pumping rate. The difference between the original water level and the water level after a period of pumping is called the *drawdown*. The *discharge rate* is determined by a flowmeter attached to the discharge pipe. The ratio between the discharge rate and the drawdown characterizes the *specific capacity* (m^3/sec) of the well. Hydrogeologists use the result of controlled pumping tests to predict the effects of future pumping on water levels.

Pumping water from a well lowers the water table around the well and creates a *cone of depression* (Fig. 7.4). Surrounding small-yield wells in productive aquifers, the cone of depression is small and shallow. A well pumped for irrigation or industrial use can withdraw so much water that the cone of depression extends for many kilometers (miles).

Locating wells too close together causes more lowering of a water table than spacing them far apart. This process is called *interference*. Interference can draw water levels so low

Box 7.3

Construction of Wells with Drilling Rigs
Either a cable-bucket rig or a rotary drilling rig is commonly used for this purpose. A cable-bucket rig churns a heavy bit up and down in pounding it through the soil or rock. A rotary rig drills its way through the strata. In either case, the hole is *cased* with a pipe to prevent borehole cave-ins. The drilling is stopped and a water pipe is lowered inside the casing when a hole has been drilled some distance below the water table. The *well point* is the lower end of the pipe with a slotted screen that allows water to enter the pipe but excludes attached sediment. Water is forced out of the well by a motor-driven submersible pump or a pump driven by a windmill in locations that lack access to electricity unless it is an artesian well.

that pumping costs will be greatly increased. In addition, legal issues can develop within a community when a high-capacity well adversely affects a neighboring well.

Management Considerations

Although the management of groundwater must be considered in the context of overall watershed management, we discuss special considerations of groundwater management here that complement the discussions on upland, riparian, and wetland watershed management in Chapters 12 and 13.

The continued use of large quantities of groundwater can create water problems. Under natural conditions, the hydrologic cycle tends to be in balance. However, continual use of groundwater resources can upset this balance. Use of groundwater resources without knowledge of the effects of continual use is unwise. In contrast, to achieve the *sustainable*

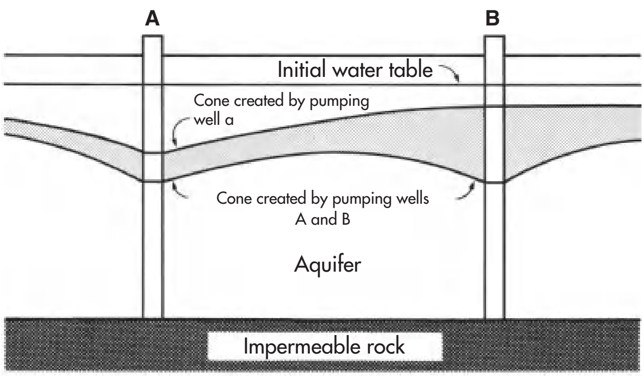

FIGURE 7.4. Cones of depression (from Baldwin and McGuinness, 1963)

use of a groundwater resource requires knowledge of basic water facts. Studies that determine the availability of groundwater resources in local areas are needed and basic research on movement and recharge is needed to make informed decisions on the sustainability of a groundwater resource. Information on groundwater quality is also necessary and methods of storing surplus water in underground reservoirs must continue to improve.

Groundwater is managed sustainably by applying the concept of *safe yield*, which refers to the annual draft of groundwater at levels that do not produce undesirable effects. For example, groundwater should not be withdrawn at rates that result in excessive lowering of the water table and, hence, incurring high pumping costs. However, a water-budget analysis of aquifers can be performed to determine the quantitative aspects of safe yield. Various inputs can be compared with outputs as follows:

$$I - O = \Delta S \tag{7.6}$$

where I is the inputs to groundwater including groundwater recharge by percolation of rainwater and snowmelt, artificial recharge through wells, and seepage from lakes and streams; O is the outputs from groundwater including pumping, seepage to lakes and streams, springs, and sometimes ET; and ΔS is the change in storage, determined as the product of change in water table elevation and specific yield for an unconfined aquifer or the product of change in potentiometric head and storativity for a confined aquifer.

Groundwater should be managed over long periods of time so that there is zero change in storage. Artificial recharge can sometimes offset pumping so that $\Delta S = O$. Depletion of groundwater storage not only affects groundwater use but also can affect surface water supplies by reducing groundwater contributions to streams, lakes, and wetlands.

The excessive withdrawal of some groundwater can have unanticipated and often detrimental effects. In karst systems, for example, sinkholes can form where groundwater withdrawal reduces water pressure to the point where sediment contained within limestone voids alters and collapses into open voids (Magner et al., 1986). Such sinkholes can become large and seriously damage property. Similarly, subsidence can occur where thick, compressible clays and silts overlie aquifers that are overly pumped. Again, the loss of groundwater pressure can lead to compaction and subsidence over large areas. This loss has been observed in Mexico City, Mexico; Bangkok, Thailand; Venice, Italy; and the southwestern USA where groundwater has been pumped excessively for decades.

Excessive groundwater pumping of groundwater near coastal areas can lead to saltwater intrusion. Reduced water pressure in freshwater aquifers creates hydraulic gradients favoring saltwater movement into the aquifer. Once an aquifer is contaminated by saltwater, the use of groundwater in the area becomes severely restricted. Costly methods of artificial recharge involving adding freshwater back into aquifers through wells can sometimes be used to reduce saltwater intrusion.

Groundwater Recharge Considerations

Several problems can result from extensive pumping of groundwater in both upland areas and the valleys below. Land subsidence is usually not of major concern in upland areas. Perhaps of greatest concern is when upland watershed inhabitants become overly dependent on groundwater resources that are not sufficient to support long-term sustained demands. Increasing human and livestock populations in remote watersheds can deplete local groundwater supplies quickly. Prolonged dry spells or droughts can then cause loss of life and serious economic losses. A point can be reached easily where digging another well or deepening existing wells are no longer viable solutions to meet water needs. Appraisals of the opportunities for water to recharge a given groundwater aquifer will be necessary in these cases.

The rate of recharge of a groundwater aquifer is governed by the availability of water to do so and the infiltration and hydraulic characteristics of the soil and rock strata in the recharge area. The time required for water to move from recharge zones to well sites can be days, months, or years. For coarse sands and fine gravels, water can travel at rates of 20–60 m/day, while the rates of travel in finer clays and dense-rock aquifers can be less than 0.001 m/day. By altering the hydraulic properties of the soil system in recharge zones, rates of infiltration and recharge can be affected. However, these impacts might not be observed at downstream well sites for long periods due to the lag time associated with travel through geologic strata. New drilling methods involving the fracturing of rock can increase the travel time of water movement.

Naturally occurring springs can provide local sources of water and are useful indicators of the location and extent of groundwater aquifers. The permeability of the aquifer and its recharge determine the discharge of a spring. The areal extent of the recharge area and its hydraulic characteristics govern the amount of recharge that takes place. As discussed earlier, springs occur where a water table intercepts the ground surface and when discharge is sufficient to flow in a small rivulet most of the time. If such flow is not evident, the resulting wet areas are called *seeps*. Springs can normally be found at the toes of hillslopes, along depressions such as stream channels, and where the ground surface intercepts an aquifer.

The dependability of springs and wells in providing high-quality water on a sustainable basis is a function of recharge, the extent of the aquifer hydraulic residence, and its yield characteristics. Perched or temporary zones of saturation that are identified by seeps or springs that flow only during the wet season can be found in upland watersheds. Extreme variability of the flow can indicate an unreliable source of groundwater or a groundwater system that has limited storage even when the flow occurs throughout the year.

The quality of groundwater can indicate whether a groundwater aquifer is part of the regional groundwater system or derived from a lake or perched strata (Box 7.4) and, by doing so, provide insight on the dependability of the aquifer and its recharge possibilities.

Effects of Vegetation on Groundwater

Much of the focus of groundwater management centers on the geological aspects of the location, volumetric extent, and hydraulic characteristics of aquifers that relate to dependability and performance of the groundwater resource. However, there are also situations where a vegetative cover can also affect groundwater. The effects of vegetation on groundwater are most often manifested in riparian communities and wetland ecosystems.

Groundwater is at or near the soil surface in most riparian and wetland ecosystems. Riparian communities consist of plants growing adjacent to streams or lakes and often have root systems in close proximity to the water table. These streamside plant communities are found in both wet and dry climates. Many riparian plant species have adapted themselves to conditions of shallow water tables or wet sites near to streams and lakes and, therefore, consume large amount of water including groundwater. Since water is available throughout their growing season, transpiration by these plants can occur at rates near potential

Box 7.4

Determining the Source of Groundwater

Specific conductance, the temperature-corrected electrical conductivity of freshwater, can be used as an indicator of the source of groundwater in certain landscapes. For example, groundwater-fed lakes and peatlands in Minnesota are rich in dissolved minerals containing calcium, magnesium, sodium, iron, and other cations. As water passes through soil, biogeochemical processes result in the accumulation of cations in the water which in turn must be charge-balanced by anions. The more cations acquired by the water, the higher the conductivity. In northeastern Minnesota, specific conductance readings greater than 120 mho/cm indicate significant contributions of water to lakes or peatlands from regional groundwater sources (Hawkinson and Verry, 1975; Clausen and Brooks, 1983). Readings less than 50 mho/cm indicate a surface water dilution, shorter subsurface residence times, and low concentrations of cations in the water. In this case, specific conductance can be used to predict the dependability of the groundwater source. Areas with calcareous-derived soils might not indicate the same distinction because surface water passing through soils, even for short periods of time, could have higher specific conductance readings due to calcite dissolution.

evapotranspiration (PET). Large quantities of groundwater can be extracted annually from underlying aquifers as a result.

Extensive riparian communities are found along ephemeral streams and on floodplains with shallow water tables in the western USA. Many of the plants in these communities are *phreatophytes*; that is, plants with extensive rooting system that can extract water directly from the water table or from the capillary fringe, therefore, impacting the sustainability of surface water resources (Poff et al., 2010; see Chapter 10 for more detail). For example, extensive stands of saltcedar (*Tamarix chinensis*) occur on floodplains throughout the southwestern USA. Following their introduction from the Old World, saltcedar has spread throughout the region to become one of the more troublesome phreatophytes. The consumption of water (ET) by saltcedar and other phreatophytes can represent a significant loss of subsurface water, including groundwater along floodplains (Table 7.3). Higher ET losses occur when the water table is shallow.

Wetlands can support of variety of vegetative communities, including trees, shrubs, mosses, grasses, and sedges that are adapted to persistent saturated soils that are often the result of shallow groundwater or groundwater discharge at the soil surface. Where wetlands are connected to groundwater, ET losses from wetland soils and plant communities can result in an annual net loss of groundwater, as annual ET can occur at PET rates. This is counter to a commonly held belief that wetlands are important source areas of water flow and important recharge areas.

TABLE 7.3. Annual estimates of evapotranspiration from phreatophytes in the southwestern USA

Species	Water table depth (m)	Annual evapotranspiration (m)	Reference
Saltcedar (*Tamarix* spp.)	1.5	2.2	Van Hylckama (1970)
	2.1	1.5	
	2.7	1.0	
Mesquite (*Prosopis* spp.)		0.3–0.5	Horton and Campbell (1974)
Cottonwood (*Populus* spp.)		1.1	Horton and Campbell (1974)

In some instances wetlands can temporarily recharge groundwater, but in many situations wetlands are net consumers of groundwater. These relationships are discussed in detail in the following section.

GROUNDWATER–SURFACE WATER EXCHANGES

The groundwater–surface water exchanges in this section focus on the linkages between groundwater and streams, lakes, and wetland ecosystems. With respect to the exchanges with lakes, our discussion centers on lakes formed as a result of lacustrine or slow-moving water or glacial processes (buried ice blocks) that result in limited vertical movement of groundwater because of fine-grained silt and clay sediment deposition. Large or riverine lakes are not relevant to this discussion because groundwater contributions are minuscule in comparison to river water flowing through riverine-formed lakes.

Linkage to Streams and Riparian Areas

Streams that flow from alpine mountain lakes typically have little to no groundwater exchange because of thin soil/alluvial sediments over bedrock. However, this type of stream is driven by unique bedrock conditions associated with mountain terrain, whereas most streams formed in alluvial sediments have a subsurface exchange component. This section discusses the common exchange components associated with streams and their riparian areas including perennial flow and the exchange of water through the hyporheic zone.

The groundwater component of a stream can be difficult to grasp because observation suggests water moving as a stream is surface water. However, in most humid parts of the world, 30% to 40% of all streamflow is derived from groundwater. How do we know groundwater becomes streamflow? How else could we explain why perennial streams keep flowing during a drought? A more difficult question is where did the groundwater enter the stream?

A linkage between a stream and groundwater takes place in the *hyporheic zone* in which saturated sediments lie below a streambed and extend laterally beneath the stream banks (Boulton, 2000). An exchange of water and materials between surface water flowing above, groundwater below, and alluvial aquifers laterally occurs within what is defined as the hyporheic zone (Fig. 7.5). This exchange is facilitated by a dynamic hydraulic gradient between the stream and adjacent groundwater aquifer that can extend outward for hundreds of meters in larger stream and river systems.

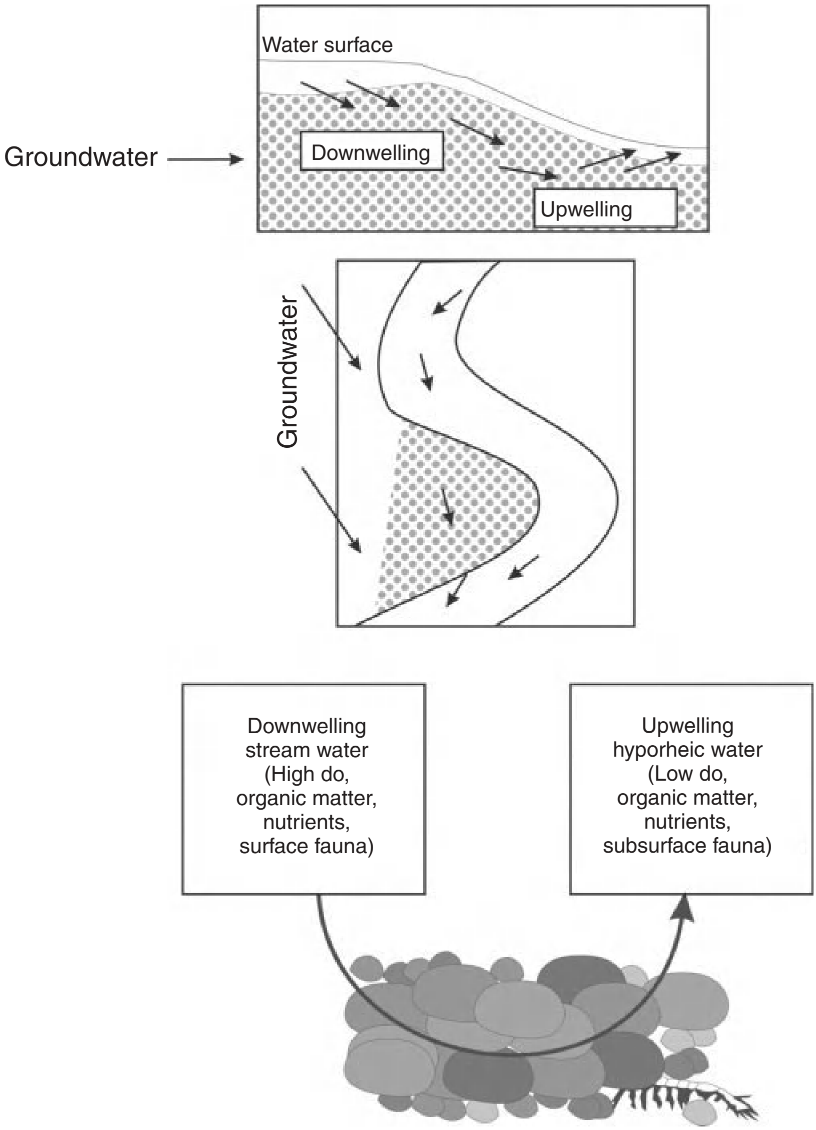

FIGURE 7.5. Water in the hyporheic zone is exchanged above with surface water, below with groundwater, and laterally with riparian corridor and alluvial aquifers (from Boulton, 2000; © Blackwell Scientific Ltd, with permission)

Because water within the hyporheic zone comes from exchanges between regional groundwater resources and the stream, its composition can vary from a mixture of groundwater and stream water to 100% of either stream or groundwater. Therefore, surface and subsurface pathways of flow are both involved with the subsurface flow occurring when the stream enters the channel bed and banks and then re-emerges downstream (Fernald et al., 2001). It is the flow of water through the hyporheic zone that promotes the biochemical processes that are important to water quality and the aquatic habitat.

Box 7.5

Sources of Baseflow to a Flowing River: One Example (Wirt and Hjalmarson, 2000)

Multiple lines of evidence have been used to identify source aquifers, quantify their respective contributions, and trace groundwater flow paths that supply baseflow to the uppermost reach of the Verde River in north-central Arizona. Groundwater discharge through springs provides the baseflow for approximately a 40-km long reach. This flowing reach is important to the increasing number of downstream water users, maintains critical habitats necessary for the recovery of native fish species, and has been designated a Wild and Scenic River because of its unique characteristics. Sources of baseflow in the reach were deduced from baseline geologic information, historical groundwater levels, long-term precipitation and streamflow records, downstream changes in baseflow measurements, analysis of water budget components, and stable-isotope geochemistry of groundwater, surface water, and springs. Taken together, this information clearly indicates that the interconnected aquifers in the headwater area are the primary sources of springs presently supplying at least 80% of the upper Verde River's baseflow. It is possible, therefore, that increased pumping of these aquifers to furnish water to the rapidly increasing human population downstream could contribute to a dewatering of this section of the river.

Groundwater that seeps into streams provides the baseflow for perennial stream systems. Therefore, if the level of the water table declines because of excessive ET losses or excessive pumping of the groundwater (Laing et al., 2009), the baseflow of these streams will also be reduced (Box 7.5). As shown by this example, surface water and groundwater are linked inextricably and, therefore, should be evaluated together (Winter et al., 1998). River-basin development such as large-scale reservoir construction for flood control, irrigation schemes, and urban expansion can affect groundwater reservoirs and vice versa. However, plans for development of a river basin frequently neglect consideration of groundwater. The rate of the natural replenishment or recharge need not limit the use of groundwater if floodwaters can be used to increase recharge. River-basin development should include an integrated program that recognizes the linkage between surface water and groundwater systems so that actions such as of flood control and artificial recharge can be coordinated.

Linkage to Lakes

Most of the water exchanges with the lakes formed as a result of lacustrine or glacial processes occur along the near-shore *littoral* margins with less exchange occurring toward the deeper portions of the lake (see Fig. 7.6). The less conductive lake bed material found

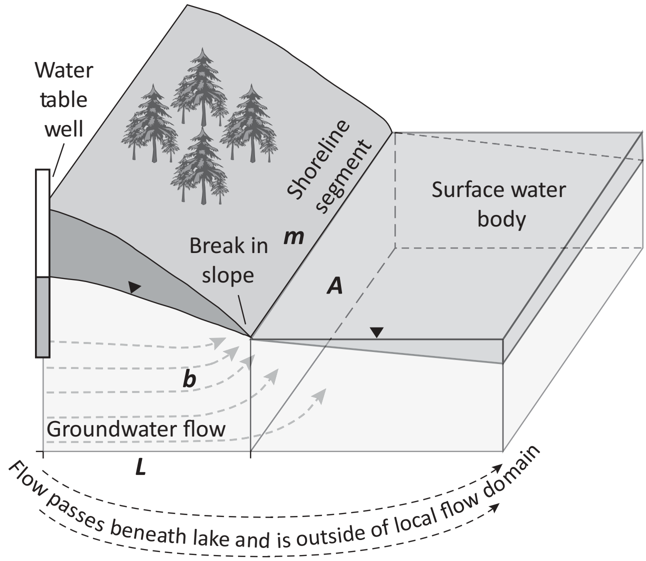

FIGURE 7.6. Typical hydraulic conditions between inflowing groundwater and a lake show how flow lines are closer to each other at shoreline and then become more distant further into the lake bed (adapted from Rosenberry and LaBaugh, 2008)

toward the deeper center portion of the lake usually remains saturated, but shallow lakes can go dry during extended droughts, allowing the cohesive lake bed sediment to crack, which in turn would allow macropore seepage. Perceived streamlines of flow into these lakes tend to compress into the most conductive strata in contact with the lake as shown in Figure 7.6.

The actual hydrologic pathways of flow into lakes are dependent on geologic and vegetative factors. The underlying geology and the processes that formed the current near-surface terrain and strata essentially drive the shape, form, and type of lake or wetland. The remnants of glacial activity have produced numerous lacustrine features across the Lake States of Minnesota, Wisconsin, and Michigan in the USA.

Vegetation during the summer can highly affect the direction of water movement (see Fig. 7.7). During the day when ET is occurring, water from both the lake and the groundwater can be extracted by plant roots and transpired into the atmosphere through plant stomata (Winter et al. 1998). In Box 7.6, the direct measurement of hydraulic head between a lake and nearby water-table wells illustrates how lake–groundwater flow directions can be determined.

Deep-water lakes will have little littoral vegetation and, as such, produce little organic sediment for microbial activity. Lakes that have 50% to 80% of their surface area defined as littoral will have varying degrees of organic soil development and associated microbial activity. The nature and character of the littoral zone defines the type of vegetation that forms around the margins of a lake. The presence or absence of phreatophytes in the lake's riparian area will in turn affect water exchange between surface water and groundwater depending on climate and season.

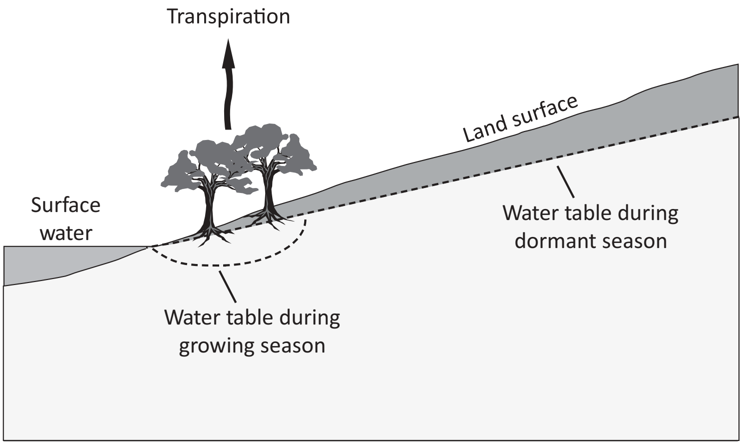

FIGURE 7.7. An example of the effect of transpiration on water movement; note the drop in the water table directly below the root zone during the growing season and the potential to pull lake water toward the root zone (adapted from Rosenberry and LaBaugh, 2008)

Box 7.6

Determining Lake–Groundwater Exchanges (Winter et al., 1998)
Williams Lake, an isolated land-locked lake in central Minnesota, started to lose water in the late 1970s in association with a regional drought. People who lived around the lake became concerned and sought government assistance. Winter began a study of Williams Lake that, in varying forms, has continued for decades and inspired the creation of a U.S. Geological Survey (USGS) research group focused on lake and wetland hydrology located in Lakewood, Colorado, USA. Measurements of hydraulic head differences in and around Williams Lake were studied to determine flow directions that explained water losses in the lake. This study informed numerous scientists across the world about lake–groundwater exchange. The Shingobee watershed, where Williams Lake is located, has been the focus of numerous USGS and peer reviewed journal articles and illustrates a classic example of a long-term sentinel study watershed. For more information on the Shingobee Headwaters Aquatic Ecosystem Project, go to http://wwwbrr.cr.usgs.gov/projects/SHAEP/

Linkage to Wetlands

The difference between a shallow lake and a wetland is defined primarily by their respective size as expressed by surface area, water depth, and by the functions collectively expressed by the vegetation, associated sediments, and microbiological activity. Wetlands are the areas inundated or saturated by surface water or groundwater at a frequency and duration sufficient to support, under normal circumstances, a prevalence of vegetation typically adapted to saturated soil conditions. Because of a persistence of saturated soil conditions, wetlands soils are anaerobic. As a result, wetlands have unique vegetation (hydrophytes) and soils (hydric soils) that distinguish them from adjacent uplands.

Wetlands occur in many landforms and climates, as streamside or riverine communities, wet meadows, depressional wetlands, peatlands, playas, saltwater marshes, forested swamplands, and coastal mangrove forests. In general, wetlands can be described as inland and coastal types (Box 7.7). In contrast to wetlands, riparian areas are the interface between terrestrial systems and transitional zones to lakes, streams, or wetlands. However, some riparian areas include wetlands. To differentiate wetlands from lakes, lakes only have vegetation along the near-shore littoral margins, while most wetlands are completely or nearly completely vegetated due to the shallow water depths. Wetlands develop organic-rich sediments that in turn provide unique habitat to specialized microorganisms that require electron transfers and the liberation of hydrogen ions to survive. Lakes by contrast are more influenced by direct precipitation, wave action, and evaporation; the water is oxic as compared to suboxic or reduced wetland water.

Coastal wetlands provide valuable food, spawning areas, and general habitat for fish that spend part or all of their lives in saltwater. These are areas that can be affected by changes in discharge, sediment, and pollutants from rivers and coastal areas; therefore, they can be affected either positively or negatively by upland watershed use and management. While the importance of coastal wetlands cannot be understated, a detailed discussion of these systems is beyond the scope of this book. Subsequent discussions of wetlands will focus on selected inland wetlands.

Inland wetlands located in the uplands, defined as *palustrine wetlands*, have mixed sources of waters and biogeochemical processes. While the names of these wetlands were originally based on the inherent vegetative and ecological functions, the sources of water and the organisms and constituents within the water also play an important role in defining the wetland features and functions. Rainfall and snowmelt runoff support some types of wetlands that are perched above regional groundwater. However, regional groundwater is typically the primary source of water for many types of wetlands. Riverine wetlands are strongly linked to groundwater discharge, but their formation is controlled by fluvial processes. Within a given valley, the slope, sediment supply, and particle-size distribution of sediments will influence the water exchange in the riparian corridor. The areas with steep valleys and channel slopes that have minimal aggraded floodplain sediment tend to be devoid of wetland features. By contrast, the areas where a tributary drops into a larger valley, where the channel slope flattens out because of sediment aggradation, create the geologic conditions for the formation of riverine wetlands. Though groundwater will flow into these areas during periods of low flow and occupy the sediment pore spaces, floodwaters will typically invade riverine wetlands during floods. The hydrologic functions and behavior of any wetland is dependent largely on the source water and the relative hydraulic residence of the source water. Chapter 13 expands this discussion on wetlands, with a

Box 7.7

Examples of Wetland Types

Inland Wetlands (Not Influenced by Saltwater)

- *Freshwater marshes*, or depressional wetlands, occur in depressions in the landscape or where changes in the surface topography result in groundwater discharge. The vegetation in these areas can be a variety of hydric plants such as cattails, reeds, grasses, sedges, and in some cases trees.
- *Peatlands* are wetlands where organic soils accumulate in the areas where either annual precipitation exceeds annual PET or where depressions concentrate water. Peatlands can be forested or nonforested, but mosses and sedges are common.
- *Riverine wetlands* occur along streams and rivers and in floodplains that are flooded periodically but where soils are saturated for a sufficient duration to form hydric soils and support wetland plants.
- *Deep-water swamps* occur as woody wetlands in many parts of the world where flooding is more pronounced and water above the soil surface is normal.
- *Fringe wetlands* occur along lakeshores or ponds.
- *Playas* are wetlands sustained by basin discharge in closed basins.
- *Southwestern cienegas* and *wet meadows* are typically found in high elevations.

Coastal Wetlands

- *Tidal salt marshes* are often dominated by salt-tolerant grasses and annual and perennial broad-leaved plants and are influenced by tidal action and high salt concentrations.
- *Mangroves* are estuarine wetlands that occur in subtropical and tropical areas and are productive forested wetlands where tide energy is lower, often found at the mouths of rivers and streams where mixing of freshwater and saltwater occurs.

more in-depth look at wetland hydrology and the role of wetlands in integrated watershed management.

Summary and Learning Points

This chapter is intended to provide the reader with a basic understanding of groundwater storage and flow characteristics, exchange linkages between surface water and groundwater, and surface water–groundwater exchanges in riparian systems and wetlands. At this point in the book, you should be able to

1. Define and illustrate a regional water table, perched groundwater, a potentiometric (piezometric) surface, a water table well, an artesian and an unconfined aquifer, a capillary fringe, a spring, and a cone of depression.
2. Explain the components of a water budget for a groundwater aquifer and contrast this water budget with a water budget for a watershed.
3. Explain the important factors that govern groundwater flow using Darcy's equation as a point of reference.
4. Describe characteristics of an aquifer that yields high quantities of groundwater on a sustainable basis.
5. Understand how surface water and groundwater exchanges occur in streams, lakes, and wetland ecosystems.
6. Describe surface water and groundwater interactions that are associated with different wetland types.
7. Describe the hydrologic and physical landscape conditions under which wetlands are formed.

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PART 2

Physical, Chemical, and Biological Linkages of Water Flow

PHOTO 4. Unstable stream channels result in bluff erosion as depicted in the stream in southern Minnesota, USA (Photograph by Mark Davidson) (For a color version of this photo, see the color plate section)

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

The hydrologic processes and pathways discussed in Part 1 of this book directly or indirectly affect soil erosion and its control, the transport of eroded soil particles into a stream channel, the accumulation and transport of sediment in the stream channel, the geomorphology of stream channels and floodplains, and the physical, chemical, and biological characteristics of water leaving a watershed. A working knowledge of erosion processes and the factors affecting erosion, sediment in streamflow, and fluvial processes is needed by the managers of watersheds along with an understanding of the effects of land-use and management practices on the quality of surface water and groundwater. Discussions of these topics are presented in the chapters comprising Part 2 of this book.

Soil erosion processes and controls are the focus of Chapter 8. Included in this chapter are discussions of surface erosion, erosion from gullies and ravines, and soil mass movement. Sediment supply, transport, and yield are the subjects of Chapter 9. Methods of measuring and describing sediment yields are also presented. Concepts of fluvial geomorphology are of importance to a hydrologist or watershed manager as are methods of evaluating and classifying valleys and stream channels within a watershed landscape. These topics are presented in Chapter 10. Physical, chemical, and biological constituents of streamflow that characterize water quality are reviewed in Chapter 11. The general quality of groundwater is also considered. Information presented in Parts 1 and 2 of this book is a requisite background to planning and implementing integrated watershed management practices – the focus of Part 3 of the book.

PHOTO 5. Soil mass erosion can provide sediment to streams as depicted in this scene in southwestern Montana, USA (Photograph by Mark Davidson) (For a color version of this photo, see the color plate section)

CHAPTER 8 Soil Erosion Processes and Control

INTRODUCTION

Soil loss from upland watersheds can occur through surface erosion, gully erosion, and soil mass movement. *Surface erosion* involves the detachment and subsequent removal of soil particles and small aggregates from land surfaces by water or wind. This type of erosion is caused by the action of raindrops, thin films of flowing water, concentrated overland flows, or the action of wind. While less serious in forested environments, surface erosion can be an important source of sediment from rangelands and agricultural croplands. *Gully erosion* is the detachment and movement of individual soil particles or large aggregates of soil in a well-defined channel. This type of erosion is a major form of geologic erosion that can be accelerated under poor land management. *Soil mass movement* refers to erosion in which cohesive masses of soil and rock materials are displaced and moved down gradient by gravity. This movement can be rapid as occurs with landslides or bluff collapse or it can be slow as with soil creep and channel slumps.

All of the above erosion processes can occur singly or in combination. People's activities such as timber harvesting, intensive livestock grazing, road construction, or row-crop agriculture can accelerate these processes. At times, it is difficult to distinguish the basic types of erosion and to determine whether they are natural processes or been accelerated by poor land-use practices. Factors that affect soil erosion and sediment movement from a watershed are summarized in Figure 8.1.

SURFACE SOIL EROSION

Surface soil erosion is a process in the physical sense that work requires the expenditure of energy. The energy is imparted to the soil surface by forces resulting from impulses

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

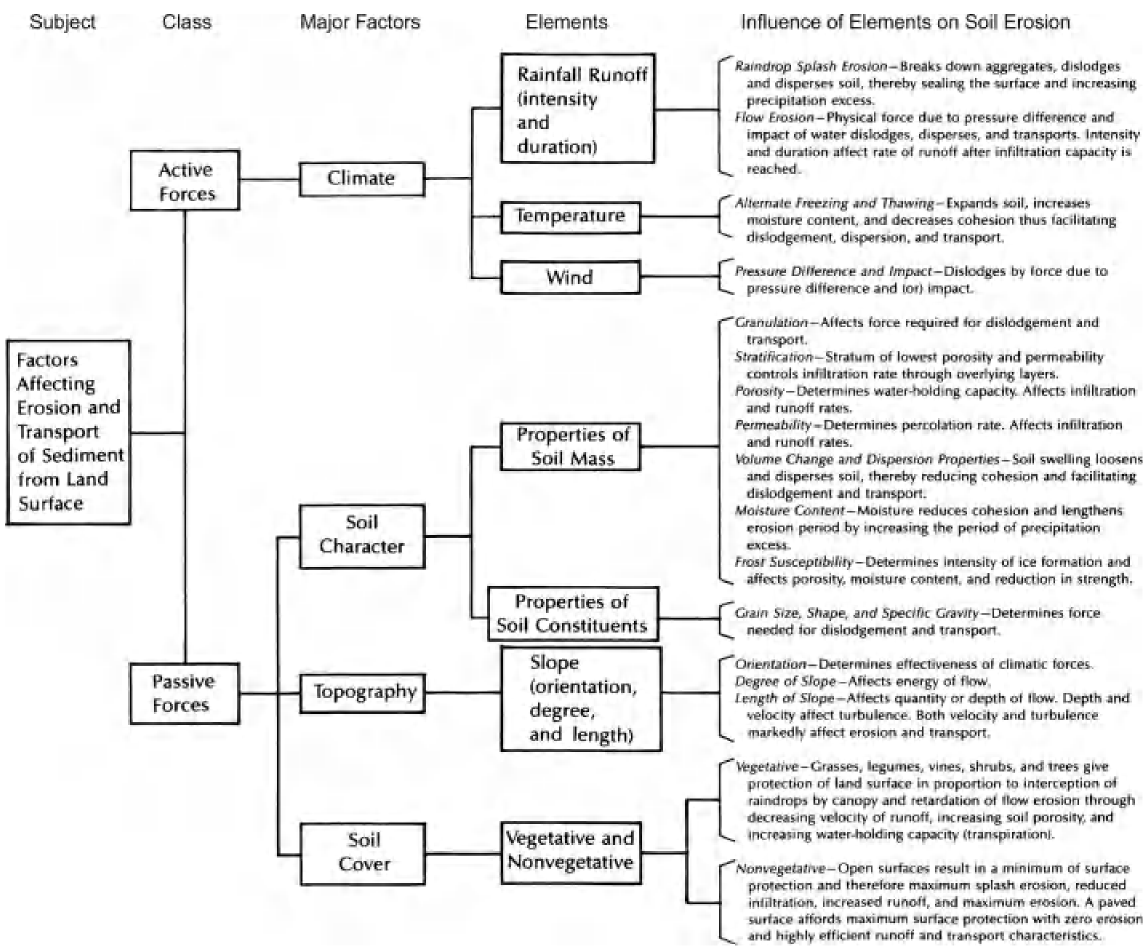


FIGURE 8.1. Major active and passive factors that affect erosion and sediment transport from a land surface (from Guy, 1964, modified from Johnson, 1961)

produced by the momentum (mass × velocity) of falling raindrops or the momentum of eddies in the turbulent flows of runoff or wind. Although wind erosion is important in some regions, we will discuss this process only briefly and then concentrate on soil erosion caused by water.

Wind Erosion

Erosion by both water and wind is a natural feature in dry regions. Wind erosion is an inevitable consequence where rainfall is inadequate to support a protective cover of vegetation. Any land use that further reduces the vegetation cover tends to accelerate wind erosion beyond that which is a natural consequence of the environment. Watersheds that support a natural vegetative cover and receive precipitation in excess of 400 mm/year experience little wind erosion. However, when soils are exposed, excessive wind erosion can even occur in regions with more than 800 mm/year annual precipitation. Wind erosion tends to diminish with increasing annual precipitation in either case.

The actions of water or wind are often complementary in their roles of removing soil particles in dry regions. For example, a soil stripped of vegetation by the abrasive action of wind-blown sand is rendered vulnerable to erosion by water. Likewise, a barren loess deposit is subject to erosion by wind. It is sometimes difficult to determine the dominant agent causing erosion on a site. Many principles of the erosion process and most of the methods of controlling the erosion apply to either process.

The largest soil particles that wind can move to any extent are about 1 mm. Very fine clay and silt particles less than 0.02 mm are lifted into the air and carried away as wind-blown dust. Sand-size particles are carried along in the air layer near the ground by saltation until they reach an obstruction where they can pile up into drifts and under extreme conditions into dunes. Just as gullies are advanced stages of water erosion, sand dunes are severe stages of wind erosion.

The erosive power of wind as that of water increases exponentially with velocity. Unlike water, however, it is not affected by the force of gravity. Therefore, slope inclination is not a factor except where sloping or hilly terrain forms barriers or influences wind direction. Similar to the effect on the erosive power of water, the length of unobstructed terrain (*fetch*) over which the wind flows is important in allowing the wind to gain momentum to increase its erosive power. Winds with velocities less than about 20 km/h at 1 m above the ground seldom impart sufficient energy at the soil surface to dislodge and put into motion sand-size particles.

Wind erosion can be controlled by planting windbreaks (shelterbelts) of trees and shrubs to reduce the velocity of blowing winds or constructing artificial but porous barriers of snow fencing or old tires. Planting of annual or perennial herbaceous species, strip cropping, or stubble-mulch tillage on agricultural fields helps to protect a soil surface from wind erosion on these lands.

Surface Erosion by Water

The dislodgement of soil particles at the soil surface by energy imparted to the surface by falling raindrops is a primary agent of erosion – particularly on soils with a sparse vegetative cover (Table 8.1). The energy released at the surface during a large storm is sufficient to splash more than 200 t of soil into the air on a single hectare of bare and loose soil. Individual soil particles can be splashed more than 0.5 m in height and 1.5 m sideways.

TABLE 8.1. Kinetic energy (K_e) associated with different intensities of rainfall and illustration of soil displacement due to rainfall impact

	Rainfall intensity (mm/h)	Kinetic[a] energy (MJ/ha/mm)[b]
Drizzle	1	0.12
Rain	15	0.22
Cloudburst	75	0.28

Source: Calculated from Dissmeyer and Foster (1980).
[a] $K_e = 1/2$ (mass)(velocity)^2.
[b] Units are megajoules per hectare millimeter.

A major impact of the impulses imparted to the soil surface by raindrops is deterioration of the soil structure by the breakdown of soil aggregates. The subsequent splashing of finer soil particles tends to puddle and close the soil surface which reduces infiltration and thereby increase surface runoff.

Surface runoff combined with the beating action of raindrops causes rills to be formed in the soil surface. *Rill erosion* is the form of surface erosion that produces the greatest amount of soil loss worldwide (Montgomery, 2007). *Sheet erosion* takes place between rills and, therefore, is also called *interrill erosion*. Sheet erosion is the movement of a semisuspended layer of soil particles over the land surface. However, minute rills are formed almost simultaneously with the first detachment and movement of particles. The constant meanders and changes in position of these small rills obscure their presence from normal observation – hence the concept of sheet erosion.

As surface runoff becomes concentrated in rills and small channels, the velocity, mass of the suspension, and the intensity of the turbulence in the flow increase downslope (Fig. 8.2). Raindrops striking the water surface add to the turbulence when the depth of runoff is shallow. Considerable energy is released by the turbulent eddies that are random in size, orientation, and velocity and provide the impulses in runoff to dislodge and entrain soil particles. The intensity of the turbulence in surface runoff is the product of velocity and depth of runoff which are both affected by the slope and the roughness of the surface over which the water flows. As the kinetic energy increases, so does the ability of the flow to dislodge and transport larger soil particles.

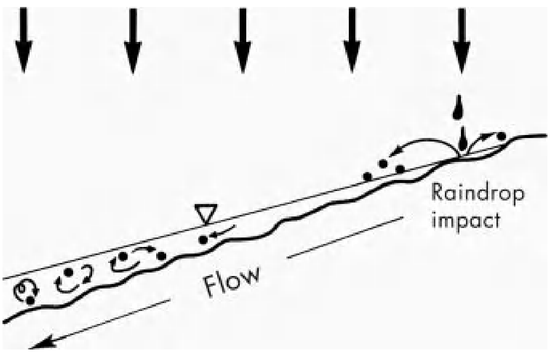


FIGURE 8.2. Surface soil erosion as a result of raindrop impact and turbulent surface runoff

In contrast to rainfall that is distributed more or less uniformly over an area, surface runoff quickly becomes concentrated in rills and channels, where its erosive power is magnified as the depth and mass become concentrated over a smaller surface area. Furthermore, as the flow picks up and carries more sediment, the abrasive action of the sediment adds to the erosive power of the runoff. On steep, unobstructed slopes and with heavy rains, soil loss in this manner can be dramatic. Such losses are also common on drylands, where normally sparse vegetative cover has been disturbed by poor land practices.

The momentum gained by surface runoff on a sloping area and the resulting amount of soil that can be lost from the area depend on both the inclination and the length of unobstructed slope. With increasing slope length, soil loss per unit length is accelerated initially but then approaches a constant rate. Soil loss increases as the inclination of the slope increases. Slope angle and slope length allow the buildup in momentum of flowing water and, therefore, are major factors in accelerating rill erosion. The steeper and longer the slope, the greater the problems of control. When rills expand and grow larger downslope, uncontrolled surface runoff is capable of creating the more spectacular gully erosion that is common on sparsely vegetated hillslopes in dryland regions. The processes of water movement on the surface are similar to the channel processes described in Chapter 9 with the primary difference being the shallow depth of rill erosion compared to channel erosion.

Measurement of Surface Erosion

Surface erosion from a small area can be measured or approximated by the use of plots, stakes, or natural landscape features such as soil pedestals. Surface erosion over larger areas such as a drainage basin can be measured by repeated reservoir surveys of designated transects or through the use of tracers (Morgan, 1995).

Erosion Plots. The most widely used method of quantifying surface erosion rates is to measure the amount of soil that washes from plots. In installing these plots, collecting troughs are sunk along the width of the bottom of the plots and walls of plastic, sheet metal, plywood, or concrete (called fabric dams) are inserted into the soil surface to form the boundaries of the plot (Fig. 8.3). The collecting trough empties into a tank or other container in which both sediment and runoff are measured. These tanks are sometimes designed with recording instruments so that the rates of flow can be measured. In other cases, the total volume of sediment and water is measured after a rainfall event has occurred.

Plots vary in size from microplots of 1 to 2 m^2 to the standard plot of 6 ft $\times$ 72.6 ft (approximately 2 m $\times$ 22 m) used for the Universal Soil Loss Equation (USLE) to be discussed later in this chapter. The techniques used and the objectives of the measurements dictate the size of the plot. For multiple comparisons of vegetation, soils, and land-use practices, microplots are less expensive and more practical than the use of rainfall simulators. Larger plots can provide more realistic estimates of erosion because they better represent the cumulative effect of increasing runoff and velocity downslope. Plots larger than the standard can yield large volumes of runoff and sediment that are difficult to store. In these cases, devices that split or sample a portion of total water and sediment flow are preferred. Such fractioning devices allow larger plots to be used to quantify the effects of larger-scale land-use practices.

FIGURE 8.3. A plot to measure runoff and soil erosion

Erosion Stakes. The insertion of stakes or pins into the soil can be used to estimate soil losses and sediment deposition that occur along hillslopes. Commonly, a long metal nail with a washer welded to the top of the nail is inserted into the soil and the distance between the head of the nail and the washer is measured. This distance increases as erosion occurs because the soil that supports the washer is washed away. If the washer causes a pedestal to form beneath the washer because of protection from rainfall impact, measurements are made from the nail head to the bottom of the pedestal. A benchmark should be established in close proximity to the stakes as a point of reference and stakes should be clearly marked so that original stakes can be accurately relocated on subsequent surveys.

Erosion stakes are usually arranged in a grid pattern along hillslopes with repeated measurements of the stakes taken over time in which the changes in soil surface are related to soil loss and deposition. This method is inexpensive compared with the plot method, but presents more difficulty in converting observations into actual soil losses in tons per hectare with measurements of bulk density.

Natural Landscape Features. Using the same principles as the stake method, erosion estimates can sometimes be made from natural landscape features. Pedestals often form beneath clumps of bunchgrass, dense shrubs, stones, or other areas protected from rainfall. As erosion removes soil from around the pedestals, the distance between the pedestal top and bottom increases. Repeated measurements of the height of residual soil pedestals provide estimates as described above. The key to applying this method is to relate measurements to a common point of reference or benchmark. Sometimes, soil that has eroded away from the base of trees can be estimated with repeated measurements of a soil surface and a point on exposed tree roots or from a nail driven into the tree trunk.

Prediction of Soil Loss

Land-use and management practices can create a variety of conditions that influence the magnitude of surface erosion. As a result, watershed managers frequently want to predict the amount of soil loss by surface erosion. Several models are available for predicting erosion caused by the action of water including the USLE, Modified Soil Loss Equation (MSLE), Revised Universal Soil Loss Equation (RUSLE), and the Water Erosion Prediction Projects (WEPP) model. The most familiar method often available for predicting soil loss is the USLE in both its original, modified, and revised forms.

Universal Soil Loss Equation. Prior to the development of the USLE, estimates of erosion rates were made from site-specific data on soil losses. As a result, these estimates were limited to particular regions and specific soils. However, the need for a more widely applicable erosion prediction technique led to the development of the USLE by the USDA Agricultural Research Service (ARS). The original USLE of 1965 was based on the analysis of 10,000 plot-years of data collected mostly on agricultural plots under natural rainfall conditions. Subsequently, erosion research has been conducted with simulated rainfall. Rainfall-simulator data are used to describe soil erodibility and provide values for the effectiveness of conservation tillage and construction practices for controlling soil erosion.

The term "universal" was given to the USLE to differentiate it from earlier erosion prediction equations that applied only to specific regions. It was applied on agricultural lands throughout the United States by 1978. The USLE was also applied to nonagricultural situations such as construction sites and undisturbed lands including forests and rangelands. However, because extensive baseline data were not available for all of these applications, a subfactor method was developed to estimate values for the C factor. The subfactor method is discussed later as the MSLE.

The English units employed in its original development are used in the following description of the USLE rather than the corresponding metric units. The basic USLE (Wischmeier and Smith, 1965, 1978) is:

$$A = RK\,(LS)\,CP \tag{8.1}$$

where A is the computed soil loss in tons/acre for the time period selected for R (usually 1 year); R is a rainfall erosivity factor for a specific area, usually expressed in terms of average erosion index (EI) units; K is a soil erodibility factor for a specific soil, expressed in tons/acre per unit of R; LS is the topographic factor, a combined dimensionless factor for slope length and slope gradient, where L is the ratio of soil loss from a given field slope length to soil loss from a 72.6 ft length under the same conditions, and S is the slope gradient factor, expressed as the ratio of soil loss from a given slope steepness to soil loss from a 9% slope under the same conditions; C is a dimensionless cropping management factor, expressed as a ratio of soil loss from the condition of interest to soil loss from tilled continuous fallow (condition under which K is determined); and P is an erosion control practice factor, expressed as a ratio of the soil loss with the practices (e.g., contouring, strip cropping, or terracing) to soil loss with farming up and down the slope.

Equation 8.1 provides an estimate of sheet and rill erosion from rainfall events on upland areas. It does not include erosion from streambanks, snowmelt runoff, or wind and it does not include eroded soil deposited at the base of slopes and at other reduced flow locations before surface runoff reaches the streams or reservoirs.

Rainfall Erosivity Factor. The rainfall erosivity factor (R) is an index that characterizes the effect of raindrop impact and rate of runoff associated with the rainstorm. It is determined by calculating the EI for a specified period of usually 1 year or one season within the year. The EI averaged over a number of these periods (n) equals R:

$$R = \frac{\sum_{i}^{n} EI_i}{n} \tag{8.2}$$

The energy of a rainstorm striking a soil surface depends on the amount of rain and all the component rainfall intensities of the storm. For any given mass in motion, the energy is proportional to velocity squared. Therefore, rainfall energy is related directly to rain intensity by the relationship:

$$E = 916 + 331\,(\log\ I_i) \tag{8.3}$$

where E is the kinetic energy per inch of rainfall (ft-tons/acre); and I is the rainfall intensity in each rainfall intensity period of the storm (in./h).

The total kinetic energy of a storm (k_e) is obtained by multiplying E by the depth in inches of rainfall in each intensity period (n), and summing:

$$k_e = \sum_{i}^{n} \left[916 + 331(\log I_t)\right] \tag{8.4}$$

The EI for an individual storm is calculated by multiplying the total kinetic energy (k_e) of the storm by the maximum amount of rain falling within 30 consecutive minutes (I_{30}), multiplying by 2 to obtain in./h, and dividing the result by 100 to convert from hundreds of ft-tons/acre to ft-tons/acre:

$$EI(storm) = \frac{2k_e I_{30}}{100} \tag{8.5}$$

The EI for a specific period (year or season) is the sum of the individual storms' EI values computed for all significant storms during that period. Usually, only storms greater than 0.5 in. are selected. The R factor is then determined as the sum of the EI values for all such storms that occurred during a 20- to 25-year period divided by the number of years (Equation 8.2).

Studies conducted on rangelands in the southwestern USA have shown that storm runoff is correlated highly with the R value for the storm. Therefore, although surface runoff could have been a parameter for inclusion in the USLE, the use of R is considered a better index for precipitation-induced erosion. Values of R in the USA vary from <20 in the high-elevation deserts of the West to >550 in the hurricane-prone Gulf Coast (Wischmeier and Smith, 1978).

Runoff events associated with snowmelt and thawing soils are common on many watersheds for which the R value must be adjusted (see references to the revised USLE discussed later).

Soil Erodibility Factor. The soil erodibility factor (K) indicates the susceptibility of soil to erosion and is expressed as soil loss per unit of area per unit of R for a unit plot. By definition, a *unit plot* is 72.6 ft long, on a uniform 9% slope, maintained in continuous fallow with tillage when necessary to break surface crusts and to control weeds. These dimensions

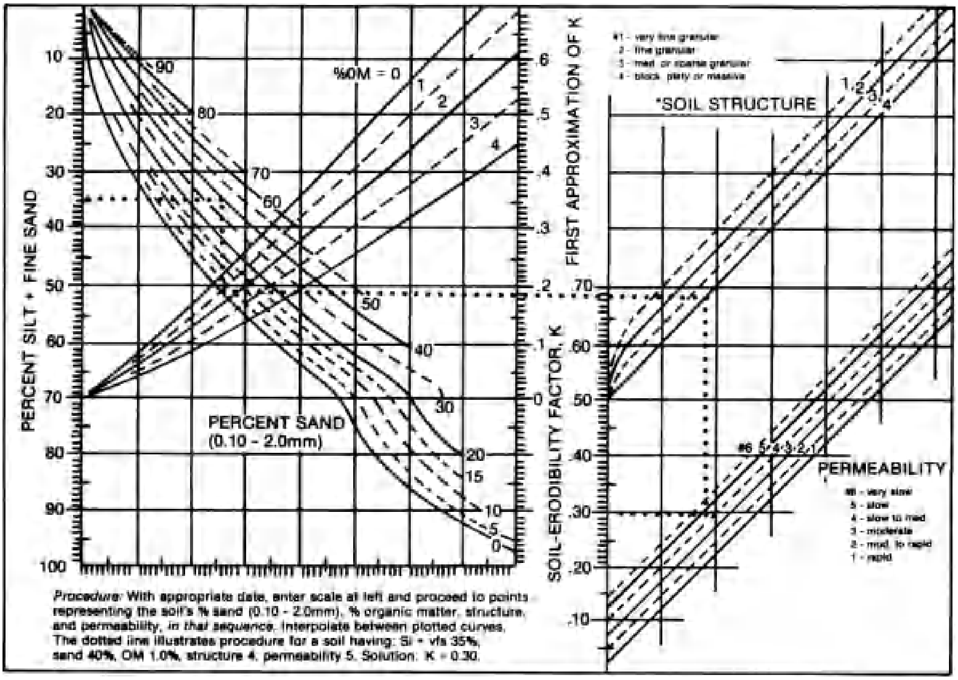


FIGURE 8.4. Nomograph for determining the soil erodibility factor (*K*) in English units (from USEPA, 1980, adapted from Wischmeier et al., 1971)

are selected because they coincide with the erosion plots used in early research in the United States. Continuous fallow is selected as a base because no single cropping system is common to all agricultural croplands and soil loss from any other plot conditions would be influenced, to a large extent, by residual and current crop and management effects both of which vary from one location to another. The K value can be determined as the slope of a regression line through the origin for source data on soil loss (A) and erosivity (R), once the ratios for L, S, C, and P have been adjusted to those of unit conditions. When the K value was originally determined with natural rainfall data, it covered a range of storm sizes and antecedent soil moisture conditions. Results of later studies conducted with rainfall simulators were used to produce a soil erodibility nomograph based on soil texture and structure (Wischmeier et al., 1971). This nomograph (Fig. 8.4) is currently used to obtain the K factor.

Slope and Gradient Factor. The topographic factors (L) and (S) indicate the effects of slope length and steepness, respectively, on erosion. Slope length refers to overland flow from where it originates to where runoff reaches a defined channel or to where deposition begins. In general, slopes are treated as uniform profiles. Maximum slope lengths are seldom longer than 600 ft or shorter than 15 to 20 ft. Selection of a slope length requires on-site judgment.

The slope length factor (L) is defined as:

$$L = \left(\frac{\lambda}{72.6}\right)^{m} \tag{8.6}$$

where λ is the field slope length (ft); and m is the exponent, affected by the interaction of slope length with gradient, soil properties, type of vegetation, etc. The exponent value ranges from 0.3 for long slopes with gradients less than 5% to 0.6 for slopes more than 10%. The average value of 0.5 is applicable to most cases.

The maximum steepness of agricultural cropland plots used to derive the S factor was 25%, which is less than many forested and rangeland watershed slopes. Recent investigations on rangelands suggest that the USLE can overestimate the effect of slope on non-crop situations. Consequently, the S factor likely will be adjusted downward in future revisions of the USLE.

The slope gradient factor (S) is defined as:

$$S = \frac{0.43 + 0.30s + 0.04s^2}{6.613} \tag{8.7}$$

where s is slope (%)

Foster and Wischmeier (1973) adapted the LS factors for use on irregular slopes; this is especially useful on wildland sites, which rarely have uniform slopes. They describe the combined factor as:

$$LS = \frac{1}{\lambda_e} \sum_{j=1}^{n} \left[\left(\frac{S_j \lambda_j^{m+1}}{(72.6)^m} - \frac{S_j \lambda_{j-1}^{m+1}}{(72.6)^m} \right) \left(\frac{10{,}000}{10{,}000 + s_j^2} \right) \right] \tag{8.8}$$

where λ_e is the overall slope length; j is the sequence number of segment from top to bottom; n is the number of segments; λ_j is the length (ft) from the top to the lower end of the jth slope segment; λ_{j-1} is the the slope length above segment j; Sj is the S factor for segment j (Equation 8.7); and s_j is the slope (%) for segment j.

For uniform slopes, LS is determined as:

$$LS = \left(\frac{\lambda_e}{72.6} \right)^m S \left(\frac{10{,}000}{10{,}000 + s^2} \right) \tag{8.9}$$

Cropping Management Factor. The cropping management factor (C) represents an integration of several factors that affect erosion including vegetative cover, plant litter, soil surface, and land management. Embedded in the term is a reflection of how intercepted raindrops that are reformed on a plant canopy affect splash erosion. Also, the binding effect of plant roots on erosion and how the properties of soil change as it lays idle are considered. Unfortunately, the manner in which grazing livestock and other plant cover manipulations change the magnitude of C is not well defined. Studies to define these cause-and-effect relationships are needed to better understand the role of the C factor in calculating annual erosion rates.

The value of C is not constant throughout the year in most cases. Although treated as an independent variable in the equation, the true value of this factor probably is dependent on all other factors. Therefore, the value of C should be established experimentally. Runoff plots and fabric dams (filter fences) are useful for this purpose. One procedure is to

- Install runoff plots on the cover complexes of interest, measure soil loss with fabric dams for each storm event, and record rainfall intensity and amount for at least a 2-year calibration period.
- Identify rainfall events with a threshold storm size great enough to produce a soil loss.

- Calculate the R value for each storm greater than the threshold storm (usually >0.5 in.) for each year of record.
- Determine the K and LS factors from the nomographs and equations given in the text.
- Solve the USLE for C for each year of record; that is, $C = A/RK(LS)$. Calculate an average value for C.

Because of the variability in storms from year to year, an average value or an expected range in R values is useful. Such estimates can be made by developing a regression relationship between the calculated R values and storm amounts measured during the calibration period. By using a historical record of daily rainfall from the nearest weather station, the R value can be determined for each storm on record and the average R for the watershed can be calculated.

In areas of the world for which there are no guidelines for the establishment of C values for field crops, it is easiest to correlate soil loss ratio with the amount of dry organic matter per unit area or the percent ground cover. C values can be estimated from published tables for permanent pastures and rangelands, idle lands, and woodlands. For forested conditions, soil consolidation, surface residue, canopy, fine roots, and residual effect of fine roots can be considered as subfactors that are used in the MSLE. The necessary tables and nomographs for estimating the C factor for nonagricultural lands are given by Dissmeyer and Foster (1985).

Erosion Control Practice Factor. The effect of erosion control practice (P) measures is considered an independent variable. Therefore, it has not been included in the cropping management factor. The soil loss ratios for erosion control practices vary with slope gradient. Practices characterized by P including strip cropping and terraces are not applicable to most forested, woodland, and rangeland watersheds. Experimental data to quantify the P factor for noncrop management practices on forested and rangeland watersheds are not available.

Applications of the USLE to a variety of conditions and geographic regions are summarized in Table 8.2. In addition, the USLE or parts thereof in original or modified forms have been incorporated into other erosion prediction technologies (Toy and Osterkamp, 1995). Consult the websites in the Webliography for this chapter for the most recent information on the USLE and current research findings.

Modified USLE (MSLE). The USLE has been modified for use in rangeland and forest environments (Wischmeier, 1975; USEPA, 1980). The cropping management (C) factor and the erosion control practice (P) factor used in the USLE have been replaced by a vegetation management (VM) factor to form the MSLE:

$$A = RL\,(LS)\,(VM) \tag{8.10}$$

where VM is the vegetation management factor, the ratio of soil loss from land managed under specified conditions of vegetative cover to that from the fallow condition on which the K factor is evaluated.

Vegetative cover and soil surface conditions of undisturbed or disturbed natural ecosystems are accounted for with the VM factor. For forests, three different kinds of effects are considered as subfactors: (1) canopy height and cover, (2) ground cover, and (3) bare

TABLE 8.2. Application of USLE and MULSE under a variety of conditions

Application	Location	Reference	Comment
Rangelands	Arizona	Osborn et al. (1977)	Evaluation of USLE for use on rangelands
	California	Singer et al. (1977)	Evaluation of USLE for use on rangelands
	Various, in western United States	Renard and Foster (1985)	Evaluate of USLE for use on rangelands (includes literature summary)
Forest lands	Manual for use of USLE on forest lands	Dissmeyer and Foster (1980)	Guidance for use of USLE under forest vegetation with adjustments to factors and subfactors
Surface mining (coal)	Alabama	Shown et al. (1982)	Equation used to evaluate erosion for assumed various phases of mining and reclamation
	Wyoming	Frickel et al. (1981)	Equation used to evaluate erosion for assumed various phases of mining and reclamation
	Pennsylvania	Khanbilvardi et al. (1983)	Equation used to estimate soil loss from interrill areas on mined and reclaimed lands
	Manual for use of USLE on surface mined lands of western United States	USDA Soil Conservation Service (1977)	Guidance for application of USLE on western mined lands
Watershed	Arizona	Fogel et al. (1977)	MUSLE as a basis for stochastic sediment yield measurements
	Mississippi	Murphree et al. (1976)	Comparison of equation estimates with sediment yield measurements
	Idaho, Colorado, Arizona	Jackson et al. (1986)	MUSLE used to estimate sediment yield from rangeland drainage basins
	Maryland	Stephens et al. (1977)	Equation used for erosion and sediment inventory of basin
	Tennessee	Dyer (1977)	Equation used in comprehensive resource planning of river basin

Source: Modified from Toy and Osterkamp (1995).

ground with fine roots. Relationships have been developed for each of the three subfactors (Fig. 8.5). The three subfactors are multiplied together to obtain the *VM* value. For grasslands where a forest canopy is not present, relationships such as the one illustrated in Figure 8.6 can be used directly. An example of an application of the MSLE is presented in Box 8.1.

More work has been carried out determining *C* values for the USLE than for *VM* factors. Therefore, there are more numerous tables of relationships for *C* than for *VM* factors. Published values of *C* can be used as a substitute for *VM* if they account for the effects described in Table 8.3.

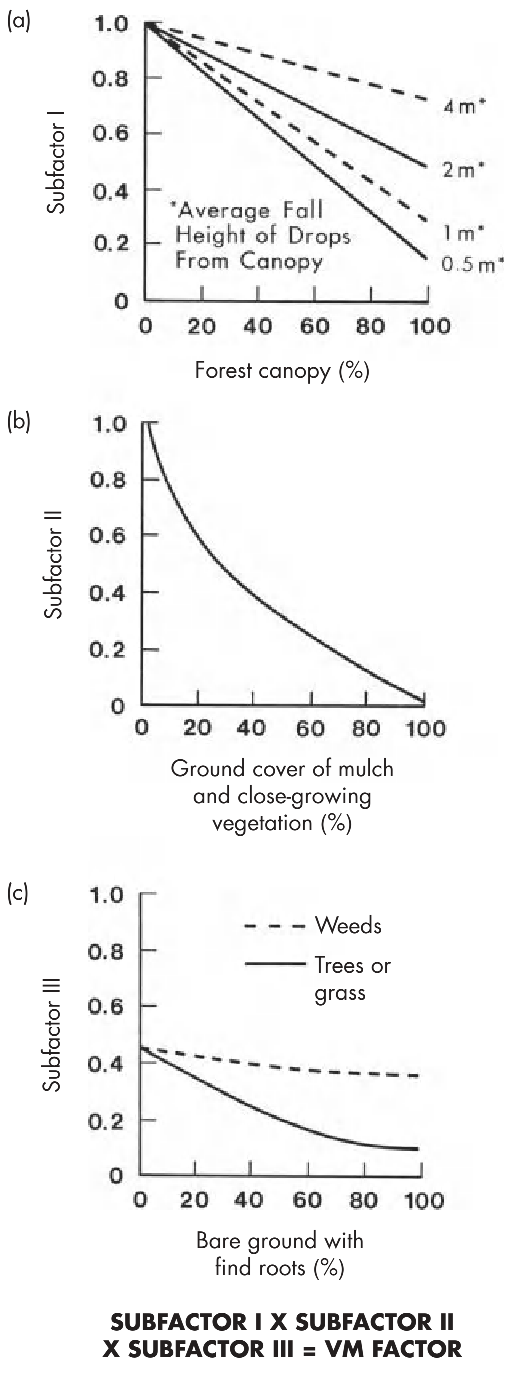


FIGURE 8.5. Relationships of forest canopy cover (a), ground cover (b), and fine roots in the topsoil (c) used to determine subfactors I, II, and III, respectively, for the VM factor (from Wischmeier, 1975)

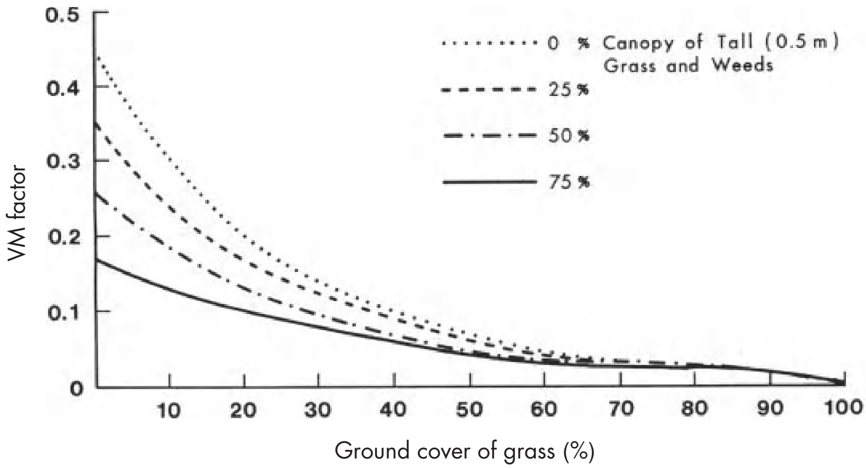


FIGURE 8.6. Relationship between ground cover conditions and the *VM* factor for the Modified Soil Loss Equation (adapted from Clyde et al., 1976)

The MSLE procedure can be used as a guide for quantifying the potential erosion on wildland watersheds under different management and land-use strategies if the principal interactions on which the equation is based are known.

Both the USLE and the MSLE require an estimate of the R factor. General estimates can be obtained for the USA from publications by the U.S. Natural Resources Conservation Service (NRCS). In locations where the R factor has not been determined, long-term rainfall-intensity records must be analyzed and R is calculated (Equations 8.2–8.5 or see websites for USLE and RUSLE).

Soil Loss Tolerance. For conservation planning, the *soil loss tolerance* needs to be established. Soil loss tolerance, also called *permissible soil loss*, is the maximum rate of soil erosion that will still permit a high level of crop productivity to be sustained economically and ecologically. Soil loss tolerance (T_e) values of 2.5–12.5 t/ha/year are often used. The numbers represent the permissible soil loss where food, forage, and fiber plants are grown. Values can be used for forest and other wildland sites but are not applicable to construction sites. Furthermore, the values do not apply for water quality standards that pertain to maximum concentrations of suspended sediment that must not be exceeded.

A single T_e value is normally assigned to each soil series. A second T_e value can be assigned to certain kinds of soil, where erosion has reduced the thickness of the effective root zone significantly to where the potential of the soil to produce biomass over an extended period of time is diminished. The following criteria are used to assign T_e values to a soil series:

- An adequate rooting depth must be maintained in the soil for plant growth.
- Soils that have significant yield reductions when the surface layer is removed by erosion are given lower T_e values than those where erosion has had little impact.

Box 8.1

Application of the Modified USLE (MSLE)

The MSLE was applied to estimating surface soil erosion before and after a watershed rehabilitation project on a 12,000 ha watershed in the Loukos Basin, Northern Morocco. This watershed is upstream of a reservoir.

Before project conditions:

Based on previous studies on rainfall and overgrazed watershed conditions in the watershed, $R = 400$, $K = 0.15$, $LS = 10$, and $VM = 0.13$ (from Fig. 8.6 based on a 30% ground cover of grass with a 25% canopy of tall weeds).

$A = R\,K\,(LS)\,(VM)$
$A = (400)(0.15)(10)(0.13) = 78$ t/ha/year (estimated annual soil loss)

The total erosion rate from the watershed was 936,000 t/year.
Post-project conditions following a watershed rehabilitation project included planting trees in the upper watershed with reseeding of grasses and managed grazing in the entire watershed were estimated as:

$^1/_2$ watershed area = 25% forest canopy with a 60% ground cover of grass,
$^1/_2$ watershed area = 80% grass cover with a 25% short-shrub cover.

These changes in vegetative cover would be expected to reduce the surface erosion (estimated by the MSLE with the modified VM factors determined from Table 8.3) as follows:

A ($^1/_2$ area) = $(400)(0.15)(10)(0.041) = 24.6$ t/ha
A ($^1/_2$ area) = $(400)(0.15)(10)(0.012) = 7.2$ t/ha

The erosion rate for the rehabilitated watershed are estimated to be 190,800 t/year. Note that these rates would only apply if grass cover was sustained through grazing management.

- For shallow soils overlying rock or other restrictive layers, little soil loss is tolerated. Therefore, the T_e should be less on shallow soils than for soils with good soil depth or for soils with underlying soil materials that can be improved by management practices.

A T_e of 11.2 t/ha/year has been used for agricultural soils in much of the USA. This maximum value was selected because soil losses in excess of 11.2 t/ha/year affect the maintenance, costs, and effectiveness of water control structures and also tend to be accompanied by gully formation in many places causing added problems for tillage operations and increasing sedimentation of ditches, streams, and waterways. Pimentel et al. (1995) estimated that erosion rates on agricultural croplands in the United States average about 17 t/ha/year and that the cost of replacing the associated loss of nutrients is \$30/ha/year. Such data suggest that T_e should be <17 t/ha/year.

Table 8.3. C or VM factors for permanent pasture, rangeland, idle land, and grazed woodland

Type and height of raised canopy[a]	Canopy cover[b] (%)	Type[c]	Cover that contacts the surface (% ground cover)					
			0	20	40	60	80	95–100
No appreciable canopy		G	0.45	0.2	0.1	0.042	0.013	0.003
		W	0.45	0.24	0.15	0.09	0.043	0.011
Canopy of tall weeds or short brush (0.5 m fall ht)	25	G	0.36	0.17	0.09	0.038	0.012	0.003
		W	0.36	0.2	0.13	0.082	0.041	0.011
	50	G	0.26	0.13	0.07	0.035	0.012	0.003
		W	0.26	0.16	0.11	0.075	0.039	0.011
	75	G	0.17	0.1	0.06	0.031	0.011	0.003
		W	0.17	0.12	0.09	0.067	0.038	0.011
Appreciable brush or bushes (2 m fall ht)	25	G	0.4	0.18	0.09	0.04	0.013	0.003
		W	0.4	0.22	0.14	0.085	0.042	0.011
	50	G	0.34	0.16	0.085	0.038	0.012	0.003
		W	0.34	0.19	0.13	0.081	0.041	0.011
	75	G	0.28	0.14	0.08	0.036	0.012	0.003
		W	0.28	0.17	0.12	0.077	0.04	0.011
Trees but no appreciable low brush (4 m fall ht)	25	G	0.42	0.19	0.1	0.041	0.013	0.003
		W	0.42	0.23	0.14	0.087	0.042	0.011
	50	G	0.39	0.18	0.09	0.04	0.013	0.003
		W	0.39	0.21	0.14	0.085	0.042	0.011
	75	G	0.36	0.17	0.09	0.039	0.012	0.003
		W	0.36	0.2	0.13	0.083	0.041	0.011

Source: From USDA Soil Conservation Service (1977).
Note: All values assume (1) random distribution of mulch or vegetation, and (2) mulch of appreciable depth where it exists. Idle land refers to land with undisturbed profiles for at least a period of three consecutive years. Also to be used for burned forest land and forest land that has been harvested less than 3 years.
[a] Average fall height of water drops from canopy to soil surface.
[b] Portion of total area surface that would be hidden from view by canopy in a vertical projection (a bird's-eye view).
[c] G, cover at surface is grass, grasslike plants, decaying compacted duff, or litter at least 2 in. deep; W, cover at surface is mostly broadleaf herbaceous plants (as weeks with little lateral-root network near the surface), and/or undecayed residue.

After having established the soil loss tolerance, the USLE or MSLE can be written as:

$$CP \text{ or } VM = T_e/[R\,K(LS)] \tag{8.11}$$

By choosing the right cropping management system and appropriate conservation practices, a value for the combined effect of C and P (or VM) that fits the equation can be established. To do so, however, it is helpful to consult the erosion index distribution curve of the area to select the most critical stages as far as rainfall erosivity is concerned.

Revised Universal Soil Loss Equation. The RUSLE predicts long-term, average-annual soil erosion by water for a broad range of farming, conservation, mining, construction sites, and other areas where mineral soil has been exposed to raindrop impacts and overland flows of water. The ARS and their cooperators initiated the development of RUSLE in the 1980s to account for temporal changes in soil erodibility and plant factors that were not originally considered in the USLE (Renard et al., 1997; Weltz et al., 1998). RUSLE also has a snowmelt-erosion component and the equation can be applied to single events. Improvements were also made to the rainfall, length, slope, and management practice factors of the original USLE model.

The RUSLE technology is computer-based and, therefore, replaces the tables, figures, and often tedious USLE calculations with keyboard entry. The improvements of the RUSLE over the USLE (Renard et al., 1997) include

- More data from different locations, for different crops and cropping systems, and for forest and rangeland erosion have been incorporated into the RUSLE.
- Corrections of errors in the USLE analysis of soil erosion have been made and gaps in the original data filled.
- The increased flexibility of the RUSLE allows for predicting soil erosion for a greater variety of ecosystems and watershed management alternatives.

RUSLE has undergone revision since its original formulation with the current version called RUSLE2, recommended for current use. RUSLE2 provides estimates of soil loss, sediment yield, and sediment characteristics from rill and interrill (sheet and rill) erosion. It uses a graphical-user interface in its operation instead of the text-based interface of earlier versions of the program. Keep in mind that RUSLE2 is not a simulation model (see Chapter 16) that attempts to mathematically replicate field processes.

The RUSLE2 computer program, a sample database, a tutorial describing the program mechanics, a set of slides that provide an overview of the operation of the program, and other supporting information are available from the ARS who developed the program, the NRCS who train perspective users of the program, and the University of Tennessee who developed the graphical-user interface. The websites for these collaborators are listed in the Webliography at the end of this chapter.

Water Erosion Prediction Project Model. The Water Erosion Prediction Project (WEPP) model is a new generation of technologies for predicting soil erosion by water that has been developed by the ARS and their cooperators. The WEPP model estimates soil erosion from single events, long-term soil loss from hillslopes, and soil detachment and deposition in small channels and impoundments within a watershed (Weltz et al., 1998). The product of the WEPP effort is a process-oriented model or family of models that are conceptually superior to the lumped RUSLE model and more versatile as to the conditions that can be evaluated.

The WEPP model operates on a daily time step, allowing the incorporation of temporal changes in soil erodibility, management practices, above- and below-ground plant biomass, litter biomass, plant height, canopy cover, and ground cover into the prediction of soil erosion on agricultural and rangeland watersheds. Linear and nonlinear slope segments and multiple soil series and plant communities on a hillslope are represented. The WEPP technology is intended to apply to all situations where erosion occurs including that resulting from rainfall, snowmelt runoff, irrigation, and ephemeral gullies.

The WEPP model is a process-based, distributed-parameter, continuous-simulation prediction model that is applicable for predicting sheet and rill erosion on hillslopes and simulating hydrologic and erosion processes on small watersheds. Included in the package that can be downloaded by a perspective user is the current version of WEPP model (v2010.1) and its operating requirements including the necessary computer software, the WEPP Windows interface, CLIGEN climate generators (versions 4.3 and 5.3), documentation, and

Box 8.2

Application of Modified USLE and the GeoWEPP Model for Predicting Soil Erosion from Roads and Trails (Renschler and Flanagan, 2008)

USA military training facilities often experience significant environmental damage from soil erosion. Much of this erosion occurs on roads and trails created by repeated military vehicle traffic during training operations. If the roads are located on steep slopes or in areas of concentrated runoff, soil loss can be large. A Geographic Information System (GIS) software package and a modified USLE were used to estimate erosion potential at Camp Atterbury located in south central Indiana. GeoWEPP (Geospatial interface to the Water Erosion Prediction Project model) was also used to estimate soil loss for the camp. Each erosion estimate was overlain with the roads and trails map. Estimated erosion levels on the camp's traffic ways with the USLE and GeoWEPP methods were then evaluated with onsite inspections of erosion conditions at Camp Atterbury. A significant correlation was found between predicted and observed erosion for both the modified USLE and GeoWEPP methods. The statistical significance for the USLE and GeoWEPP procedures allows their use in estimating erosion potential for unimproved roads and trails with confidence.

example data. A website with further detailed information on the WEPP model is found at http://ars.usda.gov/research/docs.htm?docid=10621 (accessed December 2, 2011).

An application of a geospatial interface for the WEPP model (GeoWEPP) is presented in Box 8.2.

Preventing and Controlling Surface Erosion

Avoiding erosion-susceptible situations and inappropriate land-use practices in the first place is the most economical and effective means to combat soil erosion and to maintain the productivity of watersheds. Situations that are susceptible to soil loss include

- sloping ground, particularly hills with shallow soils;
- soils with inherently low hydraulic conductivity; and
- sites where removal or denudation of vegetation is likely.

Maintaining a vegetative cover is the best means of mitigating increased soil erosion. However, a protective vegetative cover is not always present in dryland regions (Box 8.3). Instead, the need to reduce the energy of wind or flowing water suggests the use of preventive measures that are applied using simple land-use practices. The key here is to maintain the surface soil in a condition that readily accepts water. The more water that infiltrates the soil, the better the chance of reducing the erosive effects of wind and surface runoff and

Box 8.3

Surface Erosion in Dryland Africa Increases Drastically after Removal of Vegetative Cover (Harrison, 1987)

What protects Dryland Africa's vulnerable soils from the erosive actions of water and wind is vegetation. Trees, shrubs, and herbaceous plants break the force of raindrops and hold the soil in place in the face of blowing wind. With respect to water, the roots of these plants plus the activities of earthworms and termites they foster create thousands of pores and channels through which water can infiltrate into and percolate through the soil. But, when the vegetative cover is removed, soil becomes exposed to the erosive power of water and wind. The increase in surface erosion after the removal of vegetation can be spectacular. In one series of studies, the annual rate of soil loss from forests was nil – a mere 30 to 200 kg/ha. However, annual losses from agricultural croplands were nearly 90 t/ha. From bare soil, a common situation on agricultural lands at the beginning of the rainy season, the rates of annual soil loss ranged from 10 to a massive 170 t/ha. Soil that formed over hundreds of years would wash away in only one year at this latter rate of surface erosion.

sustaining plant growth. Guidelines for preventing water and wind erosion are presented in Table 8.4.

An understanding of the erosion process and the factors in the USLE and MSLE points to several measures that can be undertaken to control accelerated surface soil erosion. Actions that maintain a cover of plant material at the *soil surface* protect against the energy of rainfall impact which reduces the surface soil erosion as evident in the CP and VM relationships (see Figs. 8.5 and 8.6, Table 8.3). Plant material also increases the roughness of the soil surface, which increases the torturousity of the flow path and thereby reduces the velocity (energy) of surface runoff. Soil erodibility is also reduced by the occurrence of a network of plant roots that enhance soil strength and improve soil structure through the addition of organic matter. *ET* by the vegetation reduces soil water content between rainfall events and, as a consequence, provides additional storage for rainfall and lessens runoff.

Mechanical treatments that in effect shorten the slope length and reduce the slope inclination, lessen the energy of overland flows of water and, in doing so, reduce the quantity and velocity of surface runoff. Any actions that prevent excessive surface runoff and channelization of surface runoff will reduce the opportunity for gully formation. Strips of vegetation aligned perpendicular to the slope can also slow and reduce surface runoff.

The most effective techniques for reducing soil erosion are often those that combine several of the above actions. The loss of soil has serious implications for upland productivity but also affects the amount of sediment that can potentially reach bodies of water

Table 8.4. Guidelines for preventing water and wind erosion

Water Erosion
Avoid land-use practices that reduce infiltration capacity and soil permeability.
Encourage grass and herbaceous cover of the soil for as long as possible each year.
Locate livestock watering facilities to minimize runoff production to water bodies.
Avoid logging and heavy grazing on steep slopes.
Conduct any skidding of logs on steep slopes in upward directions to counteract drainage concentration patterns.
Lay out roads and trails so that runoff is not channelized on steep, susceptible areas.
Apply erosion control techniques on agricultural fields, and promote infiltration.
Remember that the more water that goes into the soil, the better is the chance of sustaining plant growth and reducing the erosive effects of surface runoff.
Wind Erosion
Avoid uses which will lead to the elimination of shrubs and trees over large areas.
Avoid locating livestock watering facilities on erodible soils.
Protect agricultural fields and heavy use areas with shelterbelts.
Manage animals and plants in your area to maintain a good balance between range plants, woody trees, and shrubs.
When planting shrubs and trees on grazing lands, locate and space them to reduce wind velocity.

downstream. The following discussion on soil erosion control complements the discussion presented in Chapter 12 that focuses attention on management actions – buffers and other actions to reduce sediment loads to downstream bodies of water.

Surface Erosion Control in Forest Lands

A minimal amount of surface erosion is expected in most undisturbed forest ecosystems (Fig. 8.7) with surface erosion rates rarely in excess of 0.04 t/hayear. However, activities that remove vegetative cover and expose mineral soil lead to high rates of surface erosion. Maintaining vegetative cover and litter accumulations will help to control erosion by reducing raindrop impact (and wind velocities) on a soil surface while maintaining high infiltration rates. Residual strips of vegetation alternated with clearcuts and aligned perpendicularly to the slope, function as barriers to flowing water and the downslope movement of entrained soil particles. Retaining strips of vegetation can also protect channel banks and streambeds during timber-harvesting operations (Chapter 12). Leaving residual strips of vegetation can have a minimal effect in controlling surface erosion in mountainous watersheds because of rapid channeling of surface runoff.

The nature of timber-harvesting operations can affect the magnitude of surface erosion on a watershed. Watersheds experiencing timber harvesting and associated road construction often have erosion rates in excess of 15 t/ha/year with some road construction sites exceeding 95 t/ha/year. The most serious erosion problems on forest lands are attributed to improper road and skid-trail design, location, and layout. Roads and trails are necessary for many management activities on watersheds, but their potential impact on soil erosion exceeds that of all other management activities considered.

Many potential erosion problems can be eliminated in the planning stage, before road or trail construction. A thoughtfully devised system of hauling roads will minimize the amount of mineral soil exposed. Good planning will also minimize the investment in, and

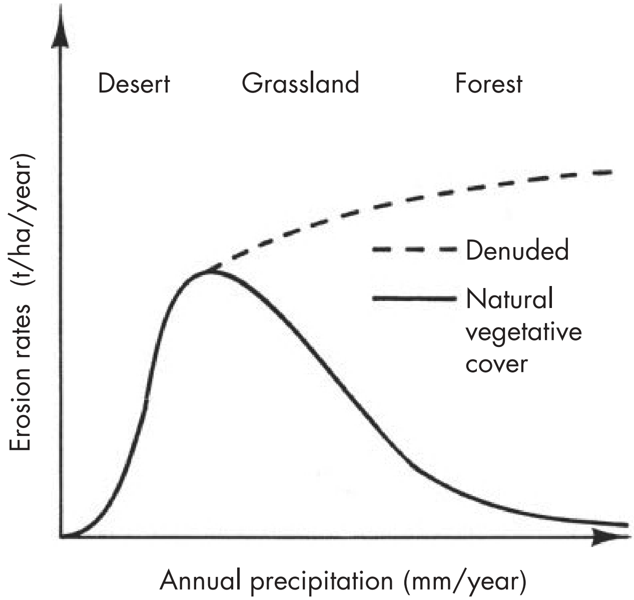


FIGURE 8.7. Relationship between erosion rates and annual precipitation for vegetation types and watershed cover conditions (modified from Hudson, 1981 and others)

maintenance of, the system. One of the more important decisions to be made in planning is the allowable width and grade of the roads as these standards will affect the area of disturbance within a watershed. Some erosion will result from roads and trails regardless of how careful the layout might be. Megahan (1977); Burroughs and King (1989), and others have identified basic principles to follow in reducing impacts of roads and road construction on erosion and sedimentation including

- Minimize the area of roads on a watershed by reducing mileage and disturbance.
- Avoid high-erosion-hazard areas when locating roads.
- Establish and maintain a vegetative cover to protect cutbanks and fill slopes along roads, on landings in timber-harvesting operations, and in other critical areas of exposed mineral soil.
- Minimize the extent of exposed soil and disturbed, unstable areas on a watershed.
- Keep roads away from stream channels to the extent possible.
- Avoid steep gradients which tend to be less stable; they expose more soil due to excessive cut-and-fill requirements, and promote high flow velocities of drainage water.
- Properly size, space, and maintain culverts to avoid road washouts.
- Provide adequate compaction of fill materials and minimize the amount of side-cast materials.

Timber-harvesting operations that have minimal effect on the compaction and disturbance of surface soils should be favored. Small cable systems can be used to remove cut trees on sites where tractors would cause excessive soil disturbance on slopes greater than 30–35%. Other log removal methods such as cable systems with intermediate supports to attain the necessary lift and extend the yarding distance can be utilized on flatter terrain. Also, double-tired low-ground-pressure vehicles with torsion suspension might replace crawler tractors. Limiting timber harvesting to the dry season or when soils are frozen in cold regions, adjusting the operations to the soil type, and minimizing the disturbance to the litter layer will lessen soil compaction and the consequent surface erosion. A variety of Best Management Practices (BMPs) are discussed in Chapter 14 as a means for reducing soil erosion–sediment loading to streams.

Surface Erosion Control on Rangelands

The greatest amount of erosion in the world occurs in dryland regions (see Fig. 8.7). These regions are generally too dry for intensive agriculture with livestock grazing often being the only economical use of the land. However, livestock grazing is uncontrolled and excessive in many countries. The establishment of appropriate range management practices must be a priority under these conditions. Otherwise, most other soil conservation practice will fail. Simply controlling livestock density and the grazing practice is often sufficient to restore depleted and eroding rangelands. The key is to maintain a healthy and extensive vegetative cover and not to reduce infiltration capacities of rangelands. If excessively grazed, these rangelands are characterized by low-plant density, compacted soils, surface runoff, and excessive erosion.

Fire is commonly used as a management tool to increase forage production of rangelands. Uncontrolled fires can reach temperatures that are high enough to reduce infiltration capacities. Wildfire can leave large areas of exposed mineral soil that are vulnerable to rainfall impact, surface runoff, and erosion. Controlled burning should not adversely impact the hydraulic properties of soils (DeBano et al., 1998).

Rangelands that are in poor condition can require reseeding of herbaceous plants and mechanical treatments to conserve water and help establish vegetative cover. Reseeding is expensive and can be justified economically only if it provides returns in forage and erosion protection over many years. Successful reseeding efforts depend on placing the correct vegetation at the correct location. Rainfall amounts and patterns, soil conditions, use of native species, proper planting methods, and grazing management following seeding must all be considered. Keep in mind that reseeding can fail with the occurrence of weather conditions such as drought and reseeding will need to be tried again.

Mechanical Methods of Controlling Surface Soil Erosion

Vegetative measures of controlling surface erosion must be accompanied by mechanical treatments when watersheds are seriously eroded. However, because of their expense, mechanical methods can be justified only if

- surface runoff and sediment from the watershed threaten important downstream developments;
- reclamation is essential to the survival of people in the area; and
- the value of the increased production of forage equals or exceeds the cost of treatment.

The purpose of mechanical treatments is to reduce surface runoff and soil loss by retaining water on a site until a vegetative cover can become established. The type of mechanical treatment chosen for implementation depends on several factors including the amount of surface runoff on the site. Often, a combination of treatments is desirable but vegetative planting and management should follow treatment as soon as possible if they are to have a lasting effect. Some of the more common mechanical methods and treatments are

1. Contour furrows – small ditches 20–30 cm deep that follow the contour-forming miniature depressions and terraces that hold the water in place until it infiltrates into the soil that reduces surface erosion and promotes plant establishment.
2. Contour trenches – large furrows that usually are required on slopes too steep for contour furrows (slopes up to 70%); they are designed to hold greater amounts of runoff and have potential for focused groundwater recharge depending on soil conditions (see Chapter 5).
3. Fallow strips – vegetation strips (about 1 m wide) along contours have proven successful on level to gently rolling land to break the slope length until desirable vegetation can become established.
4. Pitting – a technique of digging or gouging shallow depressions (20–30 cm wide to 45–60 cm long) into the soil surface to create depression storage for surface runoff and provide soil water for revegetation measures that are suited for rangeland rehabilitation, generally on gently sloping lands.
5. Basins – larger pits, usually about 2 m long, 1.8 m wide, and 15 to 20 cm deep. They store a greater amount of water and can help create pockets of lush vegetation. Basins generally are more costly to construct and are not as widely used as pitting methods.

Caution is needed when establishing contour furrows or trenches. If not placed along contours, furrows or trenches become drainage ditches that concentrate runoff and can accelerate erosion rather than prevent it. The failure of an upper furrow/trench can result in a domino effect on furrows/trenches downslope. If such a situation occurs, the resulting erosion can be much greater than that occurring in the first place.

Erosion from Gullies and Ravines

A fluvial system of which a gully is a special case is the product of a large number of interactions among many variables. These variables include the climate, relief above a base level, lithology, the area and shape of the drainage basin, hillslope morphology, soils, vegetation, human and animal activity, the slope of a channel, channel pattern and roughness, and its discharge of water and sediment. Whenever any of these variables is changed, the others also shift in response to the altered system with the exception of variables such as climate and lithology. A *gully* is a landform created by flowing water eroding sharply into soil. Gullies resemble small valleys meters to tens of meters in depth and width. Gullies are severe stages of water erosion and (in simple terms) can be thought of as channels cut into hillslopes where channels do not belong. Gully erosion occurs when the force of concentrated flowing water exceeds the resistance of the soil on which it is flowing.

Gully stabilization occurs in a natural system by moving toward a state of dynamic equilibrium when the sediment supply is balanced by the sediment transport. The stable feature is typically a *ravine* that is larger than a gully but smaller than a valley. A ravine is generally a fluvial slope landform with relatively steep incised bed and entrenched sides on the order of 20–70% in gradient. Ravines might or might not have active streams flowing along the downslope channel that originally formed them. Moreover, they are often characterized by intermittent streams since their geographic scale is usually not large enough to support a perennial watercourse. A *valley* formed by flowing water, sometimes called a *river valley*, is usually V- or U-shaped with a perennial flow. However, the shape of a valley will depend on the geology and climate. Rivers with steep gradients found in mountain ranges produce steep walls and a narrow bottom, often referred to as a *gorge*. Shallower slopes can produce broader and gentler valleys. But, in flat reaches of a river, sediment deposition occurs and the valley bottom develops a floodplain and terraces. Valleys are discussed further in Chapter 10.

Steep mountain watersheds or aeolian bluffs are prone to gully cutting. Land-use practices and poorly planned developmental activities that accelerate overland flows of water on steep hillslopes can create gullies and accelerate gully erosion processes. For example, gullies can be created on a small, poorly managed farm or by excessive livestock grazing. Eroded soils from severely gullied uplands can lead to excessive sediment supply and transport to stream channels, to lowlands where it can bury fertile bottomland soils, and to reservoirs that become filled with sediment (Trimble, 1983).

Gully Erosion Process

A *gully* is a relatively deep and recently formed channel on valley sides and floors where no well-defined channel previously existed. The flow in these channels is almost always ephemeral. Gully development can be triggered by mass wasting or tectonic movements causing a change in base level. *Base levels* are longer-term topographic elevations on a landscape that are in dynamic equilibrium with their environment. More often, gullies occur on hillslopes that have low densities of vegetation and highly erodible soils. The development of new gullies and the rapid expansion and deepening of older gullies can often be traced to removal of vegetative cover through people's activities.

A gully develops when surface runoff is concentrated at a *knickpoint*, which is where an abrupt change of elevation and slope gradient and a lack of protective vegetation occur as a result of differential rates of erosion above and below the knickpoint (Fig. 8.8). The differential rates of erosion can result from a change in the lithology of the channel. The fall of water over a knickpoint causes the bed to be undermined and the subsequent bed failure leads to migrating upgradient also called a *headcut*. Simultaneously, the force of the falling water dislodges sediment below the fall and transports it downhill, lengthening and deepening the gully in the downhill direction referred to as *downcuts*. In what is called a *discontinuous gully*, where both processes are equally important, the gully bed has a stairstep configuration. In comparison, a *continuous gully* generally gains depth rapidly from the headcut and then maintains a relatively constant gradient to the mouth, where the most active changes take place. A series of discontinuous gullies will frequently coalesce into a continuous gully (Box 8.4).

Discontinuous gullies can be found at almost any location on a hillslope (Heede, 1967; 1976). Their start is signified by an abrupt headcut. Normally, gully depth decreases rapidly downstream. A fan forms where the gully intersects the valley. Discontinuous gullies can occur singly or in a system of downslope steps in which one gully follows the next. Fusion with a tributary can form a continuous system. Gullies can also become captured by a continuous stream when shifts of streamflow on an alluvial fan divert flow from a

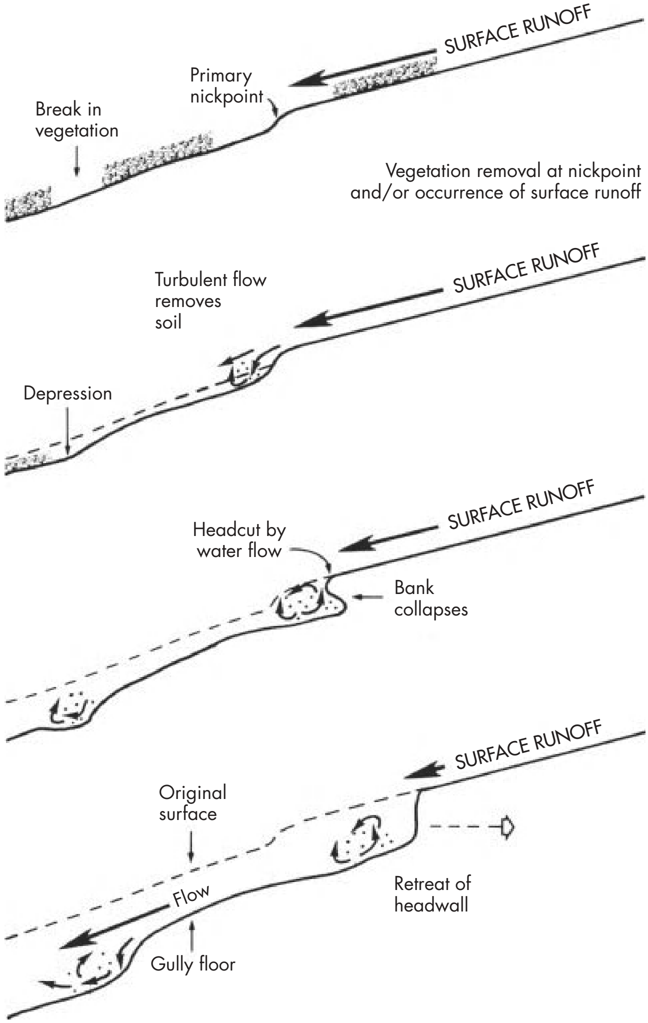


FIGURE 8.8. Illustration of gully formation and headwall retreat over time (adapted from Heede, 1967; Harvey et al., 1985)

discontinuous gully into a parallel secondary gully. At the point of overflow, a headcut that advances upstream into the discontinuous channel will develop. Here, it will form a knickpoint and intercept all flow and the gully deepens with the upstream advance of the knickpoint.

Heede (1976) describes the formation of continuous gullies as beginning with "finger-like extensions" into the headwater area. Continuous gullies usually form a gully system of stream nets. Continuous gullies are found within a diversity of vegetation types but are prominent in dryland regions.

If conditions on a watershed where gullies have formed are not improved, the gullies will continue to deepen, widen, and lengthen until a new equilibrium is reached. The process

Box 8.4

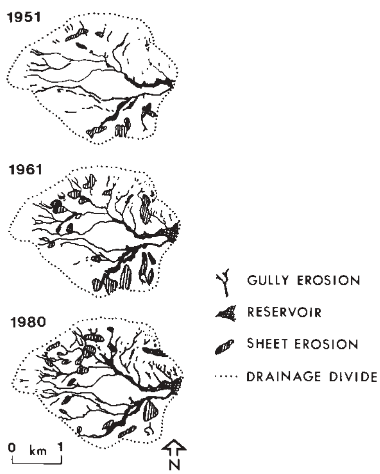
Use of Field Surveys and Aerial Photo Interpretation to Estimate Gully Erosion and Sources of Sediment (from Stromquist et al., 1985

Interpretation of aerial photos at a 1:30,000 scale and field surveys were used to identify sediment sources, transport mechanisms, and storage elements to help estimate erosion rates from a 6.15 km^2 watershed in southwestern Lesotho (see the figure below). Gully-eroded areas increased by 200,000 m^2 and produced about 300,000 t of sediment from 1951 to 1961. From 1961 to 1980, the gully-eroded areas increased by 60,000 m^2 and contributed about 90,000 t of sediment.

Sediment production at the reservoir site was estimated to be 300,000 t from gully erosion and about 80,000 t from surface erosion. The volume of sediment at the reservoir at full supply level was 267,000 m^3, which corresponds to 360,000 t of trapped sediment.

Many of the discontinuous gullies became continuous gullies from 1951 to 1961. This rapid gully expansion was explained partly by unusually high rainfall rates from 1953 to 1961, which averaged 100 mm/year more than normal. This period was preceded by an extreme drought beginning in 1944, throughout which the loss of vegetative cover made the area susceptible to erosion during the high-rainfall period of 1953–1961.

The use of aerial photography helped quantify the loss of productive area and, as used in this study, helped identify sources of sediment. In this case, only a fraction of the watershed contributed the bulk of sediments to the reservoir site.



Gully and sheet over time on a 6.15 km^2 watershed in southwestern Lesothos (Stromquist et al. 1985, © *S. African Geol. J.,* by permission).

Gully and sheet erosion over time on a 6.15 km^2 in southeastern Lesotho (Stromquist et al., 1985, © *S. African Geol. J.*, by permission).

of deposition can then begin at the gully mouth and proceed upstream until the slope of the gully sides and bottom is shallow enough to permit vegetation to become established. Plants growing along the newly formed floodplain signal the end of its active phase barring no new disturbance driving disequilibration of its downcutting and depositional forces. If the gully can be repaired such that large machinery can cross, it is considered a manageable gully. Gullies differ from ravines that are generally more established channels with deeper channels in which perennial vegetation becomes established. If channel disequilibrium occurs over time coupled with a more perennial source of water, the feature can become a small valley.

Although gully erosion is frequently the result of excessive surface runoff, subsurface flow can also dissolve, dislodge, and transport soil particles. When large subterranean voids are present in the soil, the subsurface flow can become turbulent as opposed to laminar-matrix flow and is called *pipe flow*. While common in humid, deep-soil areas (particularly in old tree-root cavities, animal burrows, etc.), pipe flow also occurs in dryland regions. The soil pipes can reach diameters of more than 1 m in some cases. Soil pipes can grow in diameter until the soil above them collapses. This process can lead to the formation of gullies, which most likely results in greater soil erosion than the actual pipe flow itself.

Gully Control

Gully erosion is the result of two main processes – downcutting and headcutting. *Downcutting* is the vertical lowering of the gully bottom and leads to gully deepening and widening, while *headcutting* is the upslope movement that extends the gully into headwater areas and increases the number of tributaries. Gully control must stabilize both the channel gradient and channel knickpoints to be effective.

Once a gully has developed, controlling gully erosion is difficult and expensive. Severely gullied landscapes can threaten on-site agricultural croplands, buildings, or road systems. Gullied watersheds can also be a source of sediment and floodwater that threaten the production of valley farmland or the effective lifetime of irrigation works and reservoirs. In cases such as these, extensive gully control projects should be undertaken. Even then, the costs of control should be weighed against probable benefits.

Permanent gully control is achieved only by returning the site to a good hydrologic condition. Obtaining this condition requires the establishment and maintenance of an adequate cover of vegetation and plant litter not only on the eroding site but also upland of the gully where runoff originates. Expensive mechanical structures in the form of check dams and lined waterways might be necessary to stabilize a gully channel temporarily to allow vegetation to become established. However, these structures require continual monitoring and maintenance over time in order to assure their initial effectiveness. The time required for vegetation to become established varies from 1 to 2 years in humid environments and 20 to 25 years in dryland regions. Therefore, the structures must be constructed accordingly. In no case should mechanical structures be considered to be permanent solutions despite how well they are constructed.

A strategy for gully control developed by Heede (1982) incorporates physical factors and parameters to establish priorities for treatment. The most critical locations in controlling gullies are at the gully head and mouth. If there is deposition at the mouth, gully widening occurs and creates an inherently unstable situation. Headcut areas are always a high priority for stabilizing a gully. Because of the high costs of controlling gullies, priority should be given to areas that yield the highest return for the least investment. Gully treatments and strategies have been published and are available from sources including publications by Food and Agriculture Organization of the United Nations (FAO); see http://www.fao.org/publications/en/ (accessed December 2, 2011).

The long- and short-term objectives of gully control must be recognized because reaching the long-term goal of revegetation directly is sometimes difficult in areas of low rainfall. Stabilization of the gully channel is the first objective of its control.

Vegetation Establishment

Where a vegetative cover can be established, channel gradients can sometimes be stabilized without resorting to mechanical or engineering measures. However, vegetation alone can rarely stabilize headcuts because of concentrated flow at these locations. Vegetation types that grow rapidly at a high-plant density with deep and dense root systems are most effective. Tall grasses that lie down on the gully bottom under flow conditions provide a smooth interface between flow and original bed and can increase flow velocities that are not suitable for gully stabilization. The higher flow velocities with such plant cover can widen the gully even though the gully bottom is protected.

Trees and shrubs can restrict high flow volumes and velocities and cause diversion against the bank. Where diverted flows are concentrated, new gullies are likely to develop and new headcuts can form where the flow re-enters the original channel. However, trees and shrubs can be planted along channel banks and on low gradients in wide gullies to form live dams that can build up sediment deposits by reducing flow velocities. Furthermore, in areas where the tree cover limits the growth of understory plants, trees can be thinned to allow for greater light penetration. The wood from the cut trees can be used to build small check dams (see below) that trap sediment.

Mechanical Structures

Gullies that are undergoing active headcutting and downcutting are difficult to stabilize and revegetate and require mechanical treatments to provide the short-term stability necessary for vegetation establishment. Once the gully gradient is stabilized, vegetation can often become established on the gully bottom. Stabilized gully bottoms will then lead to the stabilization of banks because the toe of the gully side slopes is at rest (Heede, 1976). Gully banks that are too steep for vegetation establishment can be stabilized more quickly by physical sloughing. The gully bottom should be stable, before banks are sloughed. Vegetation can be established rapidly if substantial deposits of sediment accumulate in the gully above control structures. Such deposits can store soil moisture and decrease channel gradients. The net effect of vegetation establishment in the channel and the reduced channel gradient is a decrease in peak discharge.

Mechanical or structural controls that must often be considered in controlling gullies include check dams, headcut control, and vegetation-lined waterways.

Check Dams

A *check dam* is a barrier placed in an actively eroding gully to trap sediment carried down the gully during periodic flow events (Fig. 8.9). Sediment accumulations behind a check dam function to:

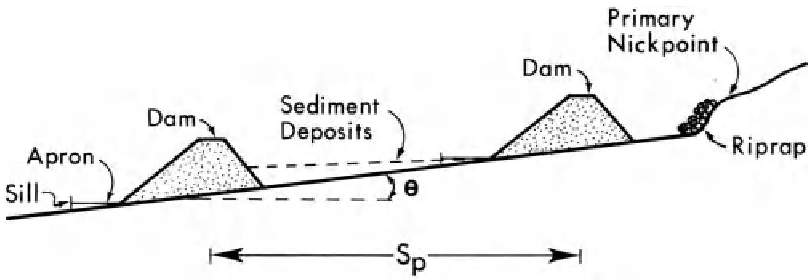


FIGURE 8.9. Diagram of placement of check dams: S_p, spacing; θ, angle of gully gradient

- develop a new channel bottom with a gentler gradient than the original gully bottom and hence reduces the velocity and the erosive force of gully flow;
- stabilize the side slopes of the gully and encourages their adjustment to their natural angle of repose, reducing further erosion of the channel banks;
- promote the establishment of vegetation on the gully slopes and bottom; and
- store soil water so that the water table can be raised, enhancing vegetative growth outside the gully.

Check dams are either nonporous or porous. *Nonporous dams*, such as those built from concrete, sheet metal, wet masonry, or earth, receive heavy impact from the hydrostatic forces of gully flow. These forces require "strong anchoring" of the dam into the gully banks, where most of the pressure is transmitted. *Earthen dams* should be used for gully control only in exceptional cases because it was likely the failure of the earth material in the first place that caused the gully. Earthen dams constructed at the mouth of gullies, often called *gully plugs*, can be effective when the upslope watershed is revegetated quickly and the storage upstream from the plug is adequate to contain the larger stormflows. The plug must have an emergency spillway to discharge water from extreme flow events that does not concentrate flow but rather spreads flow over an area stabilized by vegetation or some protective cover such as a gravel field.

Porous dams have weep holes in an otherwise impermeable structure that allows water to seep through and drain the structure, thereby creating less pressure to the banks of gullies than do nonporous dams. Since gullies generally form in erodible, soft soils, constructing porous dams is easier, cheaper, and often more effective. Loose rocks, rough stone masonry, gabions, old car tires, logs, and brush have all been used successfully to construct porous dams. Rocks are superior to most other materials for constructing inexpensive but durable porous check dams. It is important not to have large voids that allow jetting through the structure because this decreases the dam's effectiveness in trapping sediment. For loose rock dams, a range of rock sizes can be used to avoid this problem.

With the exception of gully plugs, an effective check dam should consist of three essential elements:

1. A spillway adequate to carry a selected design flow.
2. A key that anchors the structure into the bottom and sides of the gully.
3. An apron that absorbs the impact of water from the spillway and prevents undercutting of the structure.

Two additional components help to ensure the life of the structure:

1. A sill at the lower end of the apron to provide a hydraulic jump that reduces the impact of falling water on the unprotected gully bottom.
2. Protective armoring on the gully banks on the downstream side of the structure to help prevent undercutting on the sides of the dam.

The purpose of the spillway is to direct the flow of water to the center of the channel, preventing cutting around the ends of the dam. Where vegetation is slow to establish, spillways should be designed to carry the peak discharge of the 20–25 year recurrence interval peak discharge. The length of the spillway relative to the width of the gully bottom is important for the protection of the channel and the structure. Normally, spillways should be designed with a length not greater than the gully bottom to reduce splashing of water against the sides of the gully. Details of check dam designs can be found in Heede, (1976) and other references on this topic.

The stability of a check dam is increased by keying the dam into the bottom and sides of the gully to prevent water from flowing around or under the dam, making the structure ineffective. Check dams will fail if flows in the channel scour the gully side slopes below the structures, creating a gap between the dam and the bank. Turbulent flow below a check dam creates eddies that move upstream along the gully sides and erode the bank. Loose rock is effective for bank protection but should be reinforced with wire mesh secured to posts on all slopes steeper than 80–100%. Channel banks should be protected along the entire length of the apron, and banks should be protected to the height of the dam if the channel bottom is sufficiently wide.

Spacing between check dams is critical to stabilize the gully bottom and to prevent further downcutting, headcutting, and extension of the gully. Therefore, each dam should be spaced upstream at the toe of the expected sediment wedge formed by the dam below. The first dam should be constructed in the gully where downcutting does not occur, that is where sediment has been deposited at the mouth of the gully, or where the gully enters a stream system.

The spacing of subsequent dams constructed upstream from the base dam depends on the gradient of the gully floor, the gradient of the sediment wedges deposited upstream of the dams, and the effective height of the dams as measured from the gully floor to the bottom of the spillway. Heede and Mufich (1973) developed the following calculation for the spacing of check dams:

$$S_p = \frac{H_e}{K_c G \cos\theta} \tag{8.12}$$

where S_p is the spacing; H_e is the effective height of the dam, from gully bottom to spillway crest; θ is the angle corresponding to gully gradient; G is the gully gradient as a ratio ($G = \tan\theta$); and K_c is a constant, related to the gradient of the sediment deposits (S_s), which is assumed to be $(1 - K_c)\,G$.

Spacing of dams calculated by the above formula is only a guide. The choice of actual sites should be made in the field and should take into consideration local topography and other conditions such as:

1. Place the dam at a constriction in the channel rather than at a widened point if there is a choice of one or the other within a short distance of the calculated position.

2. Place the dam such that it does not receive the impact of flow of the tributary, where a tributary gully enters the main gully.
3. Place the dam below the meander where the flow in the gully has meandered within the channel.

Sediment will be deposited behind a dam on a gradient less than the gradient of the gully bottom (Fig. 8.9). The gradient of the deposit depends upon the velocity of the flows and the size of the sediment particles. The ratio of the gradient of sediment deposits to the gradient of the original gully bottom has been estimated at between 0.3 and 0.6 for sandy soils and 0.6 and 0.7 for fine-textured soils. The steeper the original gully gradient, the smaller the ratio of aggraded slope to original slope. Headcut areas above the uppermost dam should be stabilized with loose rock or riprap material.

The spacing and effective height chosen for the dams depend not only on the gradient and local conditions in the gully but also on the principal objective of the gully control. When the intention is to achieve the greatest possible deposition, the dams should have a relatively greater effective height and be spaced farther apart. If the main concern is to stabilize the gully gradient and sediment deposits are not of interest, the dams could be lower and closer together. A good rule of thumb is to keep the dam height at or below the bankfull-flow elevation (see Chapter 9).

Headcut Control

Different types of structures can be used to stabilize headcuts. All of the types should be designed with sufficient porosity to prevent excessive pressures and, therefore, eliminate the need for large structural foundations. Some type of reverse filter is also needed to promote gradual seepage from smaller to larger openings in the structure. Reverse filters can be constructed if the slope of the headcut wall is sufficient to layer material beginning with fine to coarse sand and on to fine and coarse gravel. Erosion cloth can also be effective.

Loose rock can provide effective headcut control but the flow of water through the structure must be controlled. The shape – preferably angular – and the size distribution of the rock must again be selected to avoid large openings that allow flow velocity to become too great. Care must be taken to stabilize the toe of the rock fill to prevent the fill from being eroded. Loose rock dams can dissipate energy from chuting flows and can trap sediment, which can facilitate the establishment of vegetative cover to help stabilize the toe of the rock fill.

Vegetation-Lined Waterways

The gully control measures described above are designed to reduce flow velocity within the channel and aid in the establishment of vegetation. Waterways are designed to reduce the flow in the gully by modifying the topography; to lengthen the watercourse, resulting in a gentler bed gradient; and to increase the cross section of flow, resulting in gentle channel side slopes. Shallow flows over a rough surface with a large wetted perimeter reduce the erosive power of flowing water.

A rapid establishment of vegetation lining the waterway is essential for successful erosion control. Sufficient precipitation, favorable temperature, and soil fertility are all necessary for quick plant growth. Other requisites specified by Heede (1976) include:

- the gully should not be larger than the available fill volumes;
- the valley bottom must be wide enough to accommodate a waterway that is longer than the gully;
- the soil mantle must be deep enough to permit shaping of the topography;
- the topsoil must be deep enough to permit later spreading on all disturbed areas.

Waterways are more susceptible to erosion immediately following construction than are check dams, and vegetation-lined waterways require careful attention and maintenance during the first years after construction.

Cumulative Effects of Gully Erosion

Gully formation generally results in a transport of water, soil, and chemicals from a watershed. It is important, therefore, for a watershed manager to appreciate the relationships between gully formation and land-use practices on the watershed. For example, Melton (1965) hypothesized that arroyo formation in the southwestern USA might have been promoted by decreased swale vegetation. Topographic modifications from road construction or skid trail use are a common cause for gully development in logged areas (Reid, 1993; Prosser and Soufi, 1998). Gullies often form where drainage is diverted onto unprotected slopes by roadside ditches and culverts, where ditches and ruts concentrate the flow of water or where culverts block and divert the flow over roadbeds.

Most of the changes affecting gully erosion involve altered hydrology. Surface runoff can be increased resulting in increased erosion power or channel networks can be modified which expose susceptible sites to erosion. An example of the latter is where subsurface drainage discharges into an existing stable ravine.

In assessing gully erosion and developing management solutions to control gully erosion, the cumulative effects must be considered across the watershed.

Soil Mass Movement

Soil mass movement refers to the instantaneous downslope gravity-driven movement of finite masses of soil, rock, and debris. Examples of these movements include bluffs, landslides, debris avalanches, slumps and earthflows, creep, and debris torrents (Fig. 8.10). Such movement occurs at sites where hillslopes and alterations to hillslopes experience conditions in which *shear-stress* factors become large compared with *shear-strength* factors. Because of the strong influence of gravity, these conditions are pronounced in steep, mountainous areas, particularly in humid zones that experience high-intensity rainfall events (Box 8.5) or rapid snowmelt. A general classification of hillslope failures is presented in Table 8.5.

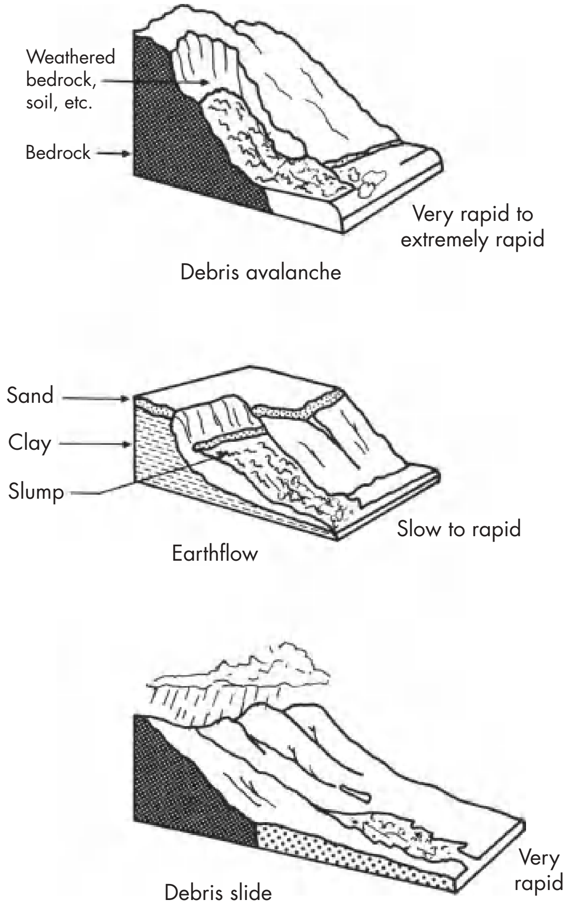


FIGURE 8.10. Illustration of soil mass movements (adapted from Varnes, 1958 and Swanston and Swanson, 1980)

Processes

The stability of soils on hillslopes is often expressed in terms of a safety factor (F). That is:

$$\mathrm{F} = \frac{\text{resistance of the soil to failure (shear strength)}}{\text{forces promoting failure (shear stress)}} \qquad \textbf{(8.13)}$$

A value of $F = 1$ indicates imminent failure; large values indicate little risk of failure. The factors affecting shear strength and shear stress are illustrated in Figure 8.11.

Shear stress increases as the inclination (slope) increases or as the weight of the soil mass increases. The presence of bedding planes and fractures in underlying bedrock can result in zones of weakness. Earthquakes or blasting for construction can augment stress. The addition of large amounts of water to the soil mantle and the removal of downslope material by undercutting for road construction, for example, are common causes of movement due to increased stress.

Box 8.5

Soil Mass Movement and Debris Flows in Taiwan

Few places in the world experience the severity of landslides and debris flows as frequently occurs in Taiwan. This country is about 75% mountainous, much of which has shallow soils overlying weak, fractured, and intensively weathered geologic formations. More than 50% of Taiwan's land areas of 35,900 km^2 has slopes steeper than 20 degrees and the island has over 100 peaks that exceed 3000 m in elevation. Population density is greater than 600 people/km^2. Every typhoon season of May to October brings torrential rainfall resulting in frequent landslides, debris torrents, and flooding of severe magnitudes. The consequent losses of life, injury, and damage to property and infrastructure are acute. People are particularly vulnerable to landslides and debris flows in the less habitable upland areas of the country, where houses are built along hillslopes and small rural communities are often clustered at the mouth of small drainages or within the floodplain itself.

Lee et al. (1990) summarized the surveys of 9900 landslides with a cumulative slide area of 16,171 ha that had occurred on 40 selected watersheds of a total area of 20,428 km^2 that occurred between 1963 and 1977. Many of these landslides occurred during the typhoon season and triggered damaging debris flows in upland drainages of these watersheds. Impacts on life and property of debris flows caused by selected typhoons in central Taiwan are summarized below to illustrate their destructive effects (Cheng et al., 1997).

Location	Dates	Rainfall event	Impacts on life and property
Tung-Men, Hualin	June 23, 1990	475 mm/3 h	29 deaths, 7 injured, 6 missing, 24 houses destroyed, severe road damage
Er-Bu-Keng, Nantou	July 31–August 1, 1996	$>$ 700 mm in less than 2 days	5 deaths, 10 houses, and 3.8 ha of fruit orchards destroyed
Tung-Fu, Nantou	July 31–August 1, 1996	$>$ 1300 mm in less than 2 days	2 deaths, 18 houses damaged or destroyed
Shen-Mu Village, Nantou	July 31–August 1, 1996	$>$ 1600 mm in less than 2 days	5 deaths, 6 injured, 8 houses destroyed, 3 ha of fruit orchards destroyed

High-intensity rainfall was the cause of all of these debris flows. The July 31 to August 1 1996 event was the result of Typhoon Herb, one of the most destructive typhoons to strike Taiwan in recent decades (see Box 1.2 for a discussion of Typhoon Herb). One of the highest-elevation rain gauges near the resulting debris flow site measured 1987 mm of rainfall in 42 hours during this typhoon event.

TABLE 8.5. Classification of hillslope failures

Kind	Description	Favored by	Cause
Falls	Movement through air; bouncing, rolling, falling; very rapid	Scarps or steep slopes, badly fractured rock, lack of retaining vegetation	Removal of support, wedging and prying, quakes, overloading
Slides (avalanches)	Material in motion not greatly deformed, movement along a plane; slow to rapid	Massive over weak zone, presence of permeable or incompetent beds, poorly cemented or unconsolidated sediments	Oversteepening, reduction of internal friction
Flows	Moves as viscous fluid (continuous internal deformation); slow to rapid	Unconsolidated material, alternate permeable, impermeable fine sediment on bedrock	Reduction of internal friction due to water content
Creep	Slow downhill movement, up to several cm per year	High daily temperature ranges, alternate rain and dry periods, frequent freeze and thaw cycles	Swaying of trees, wedging and prying, undercutting or gullying
Debris, torrents	Rapid movement of water-charged soil, rock and organic material in stream channels	Steep channels, thin layer of unconsolidated material over bedrock within channel; layered clay particles (lacustrine clays) form slippage plane when wet	High streamflow discharge, saturated soils, often triggered by debris avalanches; deforestation accelerates occurrence

Source: From Swanston and Swanson (1980).

Shear strength is determined by complex relationships between the soil and slope and the strength and structure of the underlying rock. Cohesion of soil particles and frictional resistance between the soil mass and the underlying sliding surface are major factors affecting shear strength. Frictional resistance is a function of the angle of internal friction of the soil and the effective weight of the soil mass. Pore water pressure in saturated soil tends to reduce the frictional resistance of the soil. Rock strength is affected by structural characteristics such as cleavage planes, fractures, jointing, bedding planes, and strata of weaker rocks.

Vegetation exerts a pronounced influence on many types of soil mass movement. The removal of soil water by transpiration results in lower pore water pressures, reduced chemical weathering, and reduced weight of the soil mass. Tree roots, which add to the frictional resistance of a sloping soil mass, can stabilize thin soils, generally up to 1 m in depth, by vertically anchoring into a stable substrate. Medium- to fine-root systems can provide lateral strength and also improve slope stability.

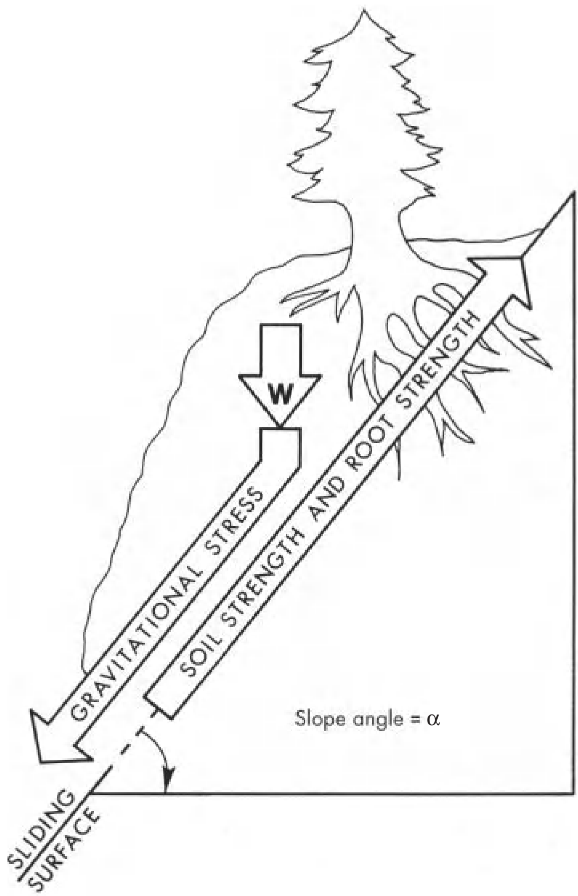


FIGURE 8.11. Simplified diagram of forces acting on a soil mass on a slope (adapted from Swanston, 1974)

Factors Affecting Hillslope Stability

Several important physical and biological factors influence slope stability and erosion (O'Loughlin, 1985; Sidle et al., 1985). Among the more important factors that act singly or in combination include:

- climate and weather factors include rainfall intensity and duration and temperature changes;
- soil factors that affect slope stability are soil strength, particle-size distribution, clay content, clay type, infiltration capacity, soil drainage condition, porosity, organic content and depth, stratification, and lithic contacts;
- physiographic features include slope steepness, slope length, and slope roughness;
- vegetative factors include cover density and type, litter thickness, root distributions, and strength of tree roots;
- water erosive forces include surface runoff and the flow of water from snowmelt;

- human-induced factors include timber harvesting and especially tree felling, clearing of forests for other land uses, road construction, and loading slopes with fill material; and
- animal factors such as overgrazing or overbrowsing by livestock or indigenous wildlife populations can influence slope stability.

Evaluating the Stability of Hillslopes

Procedures have been developed to assess soil mass movement hazards and the potential for sediment delivery to channels (Hicks and Smith, 1981; Fannin et al., 1997; Rosgen, 2006). At the landscape level, terrain evaluation procedures that utilize topographic and geologic information have been developed to provide broad categories of landslide hazard related to timber harvesting, road construction, and other management activities (Sidle, 2000). Many of these procedures are based on the factors responsible for slope stability and erosion presented above. A discussion of the methods to evaluate the stability of hillslopes is beyond the scope of this book. However, key factors that must be considered for the assessment of landslide hazard are identified in Box 8.6.

Reducing Impacts of Soil Mass Movement

Natural events such as large- and high-intensity rainfall events, earthquakes, and wildfire have a profound effect on soil mass movement. By increasing soil saturation on steep slopes that are normally unsaturated, these events can trigger landslides and debris flows on undisturbed forest lands. Wildfire can also increase the risk of landslides and debris flow. Benda and Dunne (1997) found that severe wildfire was a major cause of landslides on prehistoric landscapes in the coastal mountain ranges of Oregon. Relatively little can be done to prevent soil mass movement when natural events occur on terrain that is susceptible to slope failures.

Occurrences of landslides and debris flows can increase, however, when poorly conceived timber-harvesting activities, road construction, and vegetative conversions are carried out on these sensitive sites; these actions need not take place, however. Careful planning and implementation of these management activities help mitigate the human-induced activities compounding the occurrence of soil mass movement. Guidelines to achieve this goal are considered below in the context of cumulative effects on soil mass movement.

Cumulative Effects of Soil Mass Movement

Processes of soil mass movement attack the entire soil profile. Plant roots inhibit these processes by increasing soil cohesion. The modification of the vegetative cover, the soil system, or the inclination of a hillslope can affect soil mass movement. The impacts of land use can be estimated by relating them to factors affecting shear strength and resistance to shear. Most commonly, road construction and forest removal activities have the greatest effect on soil mass movement. Undercutting a slope and improper drainage are major factors that accelerate mass movement (Hagans and Weaver, 1987; Fannin and Rollerson, 1993). Proper road layout, design and control of drainage, and minimizing cut-and-fill (that is, earthwork) can help prevent problems. Areas that are naturally susceptible to soil mass movement should be avoided (Burroughs and King, 1989). In terms of timber-harvesting practices on steep slopes, full-suspension yarding, cable yarding, balloon logging, and other alternatives to skid roads should be used.

Box 8.6

Factors to Consider When Making Hazard Assessments of Hillslope Failure (from Swanston and Swanson, 1980)

The stability of hillslopes can be judged by evaluating the following:

Land features

- Landforms – qualitative indicator of potentially unstable land forms, e.g., fracturing and bedding planes parallel to slopes, steep U-shaped valleys.
- Slope configuration – convex or concave.
- Slope gradient.

Soil characteristics

- Present soil mass movement and rate.
- Parent material – cohesive characteristics; e.g., colluvium, tills, and pumice soils possess little cohesion.
- Occurrence of cemented, compacted, or impermeable subsoil layer – identify principal planes of failure.
- Evidence of concentrated subsurface drainage – indications of local zones of high soil moisture, springs, seeps, etc.
- Soil characteristics – depth, texture, clay mineralogy, angle of internal friction, cohesion.

Bedrock lithology and structure

- Rock type – volcanic ash, breccias, and silty sandstone are susceptible to earthflows, etc.
- Degree of weathering.
- Bedding planes or dips parallel to slope.
- Jointing and fracturing locations, directions, and relationship to slope.

Vegetative characteristics

- Root distribution and degree of root penetration in the subsoil.
- Vegetative type and distribution cover density, age, etc.

Hydrologic characteristics

- Saturated hydraulic conductivity.
- Pore water pressure.

Climate

- Precipitation occurrence and distribution.
- Temperature fluctuations, frost heaving, etc.

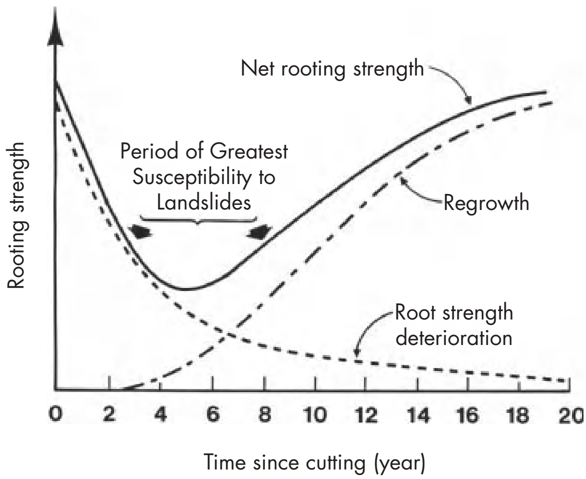


FIGURE 8.12. Hypothetical relationship of root-strength deterioration after timber harvesting and root-strength improvement with regenerating forest (from Sidle, 1985)

The removal of trees from steep slopes and particularly the permanent conversion from forest to pasture or agricultural crops can also result in accelerated mass movement. The reduced evapotranspiration with such activities can lead to wetter soils. Shear resistance is reduced by the loss and deterioration of tree roots, particularly in areas where roots penetrate and are anchored into the subsoil. For example, accelerated soil mass movement following conversion from forest to pasture has been pronounced in steep, mountainous areas of New Zealand (Trustrum et al., 1984). In many instances, therefore, maintaining tree cover on steep slopes to reduce the hazard of soil mass movement is desirable.

Routinely implemented timber-harvesting activities and regeneration practices periodically can leave hillslopes susceptible to mass movement. Root strength deteriorates rapidly as roots decay following timber harvesting (Fig. 8.12). Several years are required before the regrowing forest exhibits root strength that is equivalent to that of mature forests in the Pacific Northwest (Sidle, 1985, 1991). As a consequence, there is generally a 3- to 8-year period when net root strength is at a minimum. However, the first several years after timber harvesting also coincide with the period of maximum water yield increases caused by reduced evapotranspiration (see Chapter 3). The result is more frequent occurrences of shallow landslides on steep slopes for several years following logging. The period of susceptibility is somewhat species dependent; that is, it is affected by the rate of root decay and the rate of regrowth of the tree species being managed.

SUMMARY AND LEARNING POINTS

Erosion of the soil by water can occur as surface erosion, gully erosion, and soil mass movement from upland watersheds and hillslopes. After completing this chapter you should have gained an understanding of soil-surface erosion, gully erosion, and soil mass movement and how these processes can reduce the productive capacity of a watershed and affect sediment loading to downstream channels. More specifically, you should be able to

1. Describe the process of detachment of soil particles by raindrops and transport by surface runoff. How does this differ for wind erosion?
2. Explain how land-use practices and changes in vegetative cover influence the processes of soil detachment described above.
3. Understand how to apply the USLE and the MSLE (or RUSLE) in estimating soil erosion under different land-use and vegetative-cover conditions.
4. Explain how surface erosion can be controlled, or at least maintained at acceptable levels in different landscapes and vegetative types.
5. Understand differences between a gully, ravine, and valley and how gullies are formed.
6. Explain the role of structural and vegetative measures in controlling gully erosion.
7. Describe the different types of soil mass movement and explain the causes of each.
8. Explain shear stress and shear strength as they pertain to soil mass movement.
9. Explain how different land-use impacts, including road construction, timber harvesting, and conversion from deep-rooted to shallow-rooted plants, affect both gully erosion and soil mass movement.

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CHAPTER 9 *Sediment Supply, Transport, and Yield*

INTRODUCTION

The amount of sediment in a stream, river, lake, or reservoir is of interest in watershed management. Excessive sediment supply in streamflow can be indicative of undesirable soil erosion from varying sources within a watershed including the uplands, hillslopes, and unstable stream channels. Excessive sediment can adversely affect water quality and aquatic habitat and is one of the primary targets of the Total Maximum Daily Load (TMDL) provisions of the Clean Water Act (CWA) in the United States aimed at reducing nonpoint pollution from silvicultural and agricultural practices (see Chapter 14). Excessive sediment deposition can also adversely impact on the stability of stream channels, water conveyance systems and associated infrastructures, and reservoir storage. Excessive sediment deposition in a reservoir can reduce the economic life of the reservoir. Similarly, high levels of suspended sediment can damage turbines that provide hydro-electrical power. As a consequence, there are both environmental and financial costs of excessive sediment supply in water bodies.

Determining what constitutes excess sediment in a stream is not a straightforward process, however. It is necessary to recognize that soil erosion and sedimentation processes occur naturally and that the levels of sediment in streamflow can vary by region and from one time period to another. Relationships between erosion sources and downstream sedimentation involve the dynamics discussed in the previous chapter and complex channel processes related to sediment supply, transport, and fluvial mechanics. A comprehensive treatment of this subject is beyond the scope of this book. However, the basic processes affecting sediment supply, transport, and deposition as related to valley and channel dynamics will be examined. Methods of measurement and analysis of sediment data are also

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described. Watershed management practices and land-use activities that directly and indirectly affect sedimentation and stream channels are discussed here and in the context of integrated watershed management in Chapter 12.

Sediment Supply and Transport

Sources of Sediment

Sediment is the product of soil erosion, whether it occurred as upland surface erosion, hillslope and gully erosion, soil mass movement, bluff erosion or streambank, and bed erosion. The amount of sediment contributed to a stream channel from the surrounding watershed is dependent on many factors including proximity of the erosional features to the channel, the shear forces acting upon soil and rock, the characteristics of sediment particles, and the efficiency by which sediment particles are transferred from one part of the landscape to another. Much of the sediment is deposited at the base of hillslopes, in riparian buffer strips (see Chapter 13), on terraces, on floodplains following high flood events, or within active river channels. Only a portion of sediment is passed through and out of a watershed during storm events. A watershed that has adjusted to its unique climate, geology, and land-use activities will have a “balanced amount” of sediment supplied to its channel(s). However, some watersheds require prolonged periods of time to reach a quasi-dynamic equilibrium because of the nature of the geologic material or geologic activity, climate change, or land-use changes that increase or decrease streamflow discharge.

While soil erosion occurring on a watershed can be determined (see Chapter 8), it is difficult to determine how much of the eroded material will be transported to and deposited on a specified location over time. The following section focuses attention on the processes of sediment transport in the channel–that is, after eroded material reaches the channel.

Energy Relationships of Streams

We introduce energy concepts related to the flow and resistance to flow of unbound water and sediment in a channel at this point in the chapter. Water flowing from a watershed is moving from a state of higher energy status to one of lower energy status, the base level being sea level. Evaporation, wind, and precipitation redistribute water and can return it back to the watershed at an elevated energy status by virtue of its elevation above sea level (see Chapter 2). In responses to gravity, much of the water in the form of excess precipitation is returned to the oceans and the energy status is reduced through the process of channel flow.

Once water reaches a stream channel, the rate and type of flow are determined by gravity and resistance forces of friction (recall Manning’s “*n*” discussed in Chapter 6). Gravity forces are expressed as a continuous energy gradient called the *hydraulic gradient*. The hydraulic gradient is a potential energy gradient that exists because of the elevation of water above sea level. Although the overall hydraulic gradient is determined by the change in elevation from the highest to the lowest elevations in the watershed, the gradient is not uniform throughout a stream channel. Hydraulic gradients are generally steepest at the upper most part of the watershed and diminish as the channel nears sea level.

Water that is ponded in the upper portions of a watershed has a high potential energy but cannot perform work until it is released. Water drains from soil and reaches a stream channel under the force of gravity and flows from its higher energy state to a lower energy state downstream. The stream's potential energy is converted to kinetic energy by the velocity of the flowing water in this process (see Equation 2.14 in Chapter 2).

Water flow in a channel is governed by energy relationships on the basis of the Bernoulli equation that states:

$$(P/\rho g) + (V^2/2g) + z = \text{constant} \quad \textbf{(9.1)}$$

where P is the pressure (in units of bars or newtons per square meter); ρ is the density of fluid (kg/m^3); g is the acceleration due to gravity (m/s^2); V is the velocity (m/s^1); and z is the elevation above some datum (m).

The three components in Equation 9.1 have units of length (m) and can be considered as pressure head, velocity head, and elevational head, respectively.

For a given streamflow discharge value (Q), we know from the conservation of mass principle that even though channel dimensions can change from one section to another, the products A_1V_1 at section 1 = A_2V_2 at section 2. Therefore, if we consider the first term in Equation 9.1 to be equivalent to water depth in the channel, the velocity and depth of flow will change for a specified streamflow discharge in response to changing channel dimensions of width and bottom configuration. The overall change in the energy status of a stream, therefore, can be accounted for by the change in the water surface elevation of the stream, that is, the slope.

The Bernoulli equation is a one-dimensional energy equation that illustrates the above relationships for a section of stream channel from its upstream location (subscript 1) to its downstream location (subscript 2):

$$z_1 + D_1 + (V_1^2/2g) = z_2 + D_2 + (V_2^2/2g) + h_\mathrm{L} \quad \textbf{(9.2)}$$

where D is the mean water depth (m); and h_L is the head loss due to energy losses associated largely with friction.

The specific energy (E_s) for a given channel section with a small slope and a selected streamflow discharge is a function of water depth:

$$E_\mathrm{s} = D + (V^2/2g) \quad \textbf{(9.3)}$$

These energy relationships have important effects on fluvial processes and are the basis for several of the relationships and terms defined in Table 9.1.

Subcritical and Supercritical Flow

The *Froude number* (*Fr*) is a dimensionless parameter that is used as a quantitative measure of whether subcritical, critical, or supercritical flow will occur. This number is calculated by:

$$F_r = V/(gD)^{1/2} \quad \textbf{(9.4)}$$

where V is the average velocity in the cross section of measurement; g is the acceleration due to gravity; and D is the average water depth.

If $F_r < 1$, subcritical flow occurs, critical flow occurs when $F_r = 1$, and supercritical flow occurs when $Fr > 1$. At and below critical flow, a more or less stable relationship exists between a given depth of flow and the ensuing rate of streamflow. It is at or near critical flow, therefore, that the "best measurements" of streamflow are generally obtained. Such relationships are often used in the design of weirs and flumes. *Subcritical flow* is tranquil and

TABLE 9.1. Terminology used in describing channel dynamics and processes

Term	Definition
Uniform flow	Flow conditions in which the water surface is parallel to the streambed; depth and velocity of flow are constant over the channel reach.
Varied flow	In contrast to uniform flow, flow velocity and depth change over the channel reach.
Steady flow	Depth and velocity of flow do not change over a given time interval at a point within the stream channel.
Unsteady flow	Conditions of flow in which depth and velocity change with time at a point within the stream channel (e.g., when a wave passes past a point in the channel).
Laminar flow	Flow of fluid elements in a stream in parallel layers past each other in the same direction but at different velocities.
Turbulent flow	Flow of fluid elements in a stream past each other in all directions with random velocities (e.g., eddies).
Critical flow	Occurs when the Froude number = 1 (see Equation 9.4), which is a condition of minimum specific energy (E_s) for a given discharge and channel condition.
Critical depth (D_c)	For any given channel section and discharge, this is a depth of water that corresponds to the minimum specific energy (E_s); if water depth $>D_c$, the flow is *subcritical*, if water depth $<D_c$, the flow is *supercritical*.
Aggradation	A process by which sediments collect in streambeds, floodplains, and other water bodies and become deposited, raising their elevations.
Degradation	Process by which stream beds and flood plains are lowered by erosion and removal of material from the site.
Dynamic equilibrium	A channel system that is sufficiently stable so that compensating changes (either aggradation or degradation) will not alter the equilibrium.

exerts relatively low energy and associated shear stress on the channel banks and beds. In contrast, *supercritical flow* has high energy and, as a consequence, can damage unprotected stream channels through considerable shear stress on the bed and sides of channels.

Laminar and Turbulent Flow

Streamflow can also be characterized by the movement of individual fluid elements with respect to each other. This movement results in either laminar or turbulent flow. Each fluid element moves in a straight line with uniform velocity in *laminar flow*. There is little mixing between the layers or elements of flow and, therefore, no turbulence. Conversely, *turbulent flow* has a complicated pattern of eddies that produce random velocity fluctuations in multiple directions. Changes in flow lines during turbulent flow can lead to specific vectors of flow against streambanks and structures, increasing shear stress. Turbulent flow is the normal condition in streams. The *Reynolds number* is a dimensionless measure used to distinguish quantitatively between laminar and turbulent flow. The Reynolds number (R_e) is determined by:

$$R_e = \left(\frac{V\,D}{V}\right) = \left(\frac{inertial\ force}{viscous\ force}\right) \tag{9.5}$$

where V is the average velocity in the cross section of measurement (m or ft/s); D is the average water depth (m or ft); and v is the kinematic viscosity (m^2 or ft^2/s).

A Reynolds number less than 2000 normally indicates laminar flow and more than 2000 turbulent flow in natural stream channels. These numbers are related the energy gradient or slope and streamflow discharge. Typically, a steeper slope over a short distance; for example, a riffle will produce turbulent flow. More information about the channel cross-sectional area is required to define the nature of the flow, however.

Sediment Movement

Both water and sediment comprise the flow of a stream. As water and sediment move from a higher to a lower energy state, energy must be released. The stream dissipates energy as heat from friction and by performing work on the channel and sediments. It is this process of work by the flow of water and sediment that forms the stream channel and changes the slope of the channel.

Channel erosion is not the only way that a stream can dissipate excess energy. Energy can also be internally dissipated through turbulent flow–the eddies and boils in a stream. When a stream loses energy to its channel, momentum is transferred from the flowing water to the channel; this change in momentum is a *force*. When sufficient force is applied to the channel, sediment will be eroded and carried away. The energy that is lost in a section of channel (h_L) is a function of channel roughness (n), the hydraulic radius (R_h), and velocity (V) of the stream as shown below:

$$h_L = n\left(\frac{1}{4R_h}\right)\left(\frac{V^2}{2g}\right) \tag{9.6}$$

As discussed earlier in Chapter 6, channel roughness includes channel vegetation, woody debris, bottom forms, and the morphological features of the channel.

The resisting force exerted by the channel is a shear stress (τ), referred to here as *shear resistance*. The shear resistance is that shear stress generated on the bed and banks or the wetted perimeter of a channel in the direction of flow. Expressed as per unit of wetted area, shear resistance (τ_c) can be defined as:

$$t_c = \gamma(R_h)\, s_b \tag{9.7}$$

where γ is the specific weight of the fluid and s_b is the channel bed slope.

Channel continuity (ε) captures the balance between erosional shear stress and resistance to flow or shear strength. Channel continuity is the quasi-dynamic equilibrium between opposing forces expressed in Equation 9.8.

$$\varepsilon = k\,(\tau_o - \tau_c) \tag{9.8}$$

where (τ_c) is the boundary shear stress applied to the bed and bank material and k is the erodibility coefficient.

Types of Sediment Transported in a Stream

The *sediment discharge* of a stream is the mass rate of transport through a specified cross section of the stream and is usually measured in milligrams per liter (mg/L) or parts per million (ppm). Sediment discharge contains fine particles that are transported in suspension.

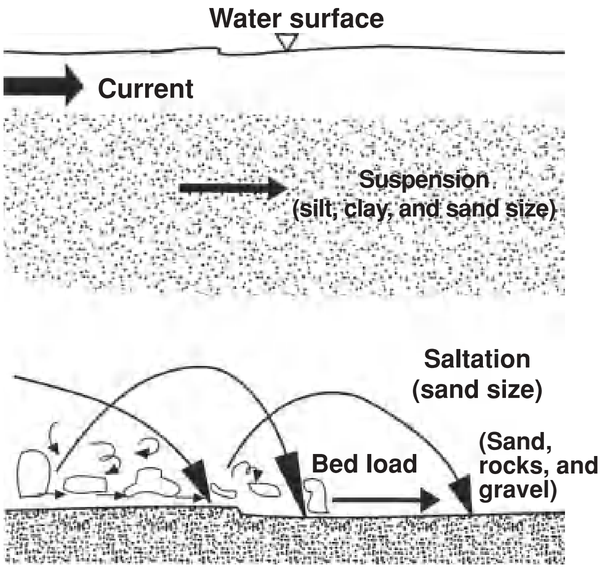


FIGURE 9.1. Transportation of particles of sediment in running water

Part of the *suspended load* is called the *wash load*. The wash load is made up only of silt and clay, while suspended sediment also includes fine sand-sized particles. The *bed load* consists of coarse sand, gravel, or larger cobbles and is transported along the stream bottom by traction, rolling, sliding, or saltation (Fig. 9.1). Sediment particles are moved when eddies formed by turbulent flow dissipate part of their kinetic energy into mechanical work.

Suspended Load. Particles can be transported as a suspended load if their *settling velocity* is less than the buoyant velocity of the turbulent eddies and vortices of the water. Settling velocity primarily depends upon the size and density of the particle. In general, the settling velocity of particles less than 0.1 mm in diameter is proportional to the square of the particle diameter, while the settling velocity of particles greater than 0.1 mm is proportional to the square root of the particle diameter. Once particles are in suspension, however, little energy is needed for transport. A "heavy" suspended load decreases turbulence and makes the stream more efficient. Concentrations are highest in shallow streams where velocities are high.

The concentration of sediment in a stream is lowest near the water surface and increases with depth. Silt and clay particles less than 0.005 mm in diameter are generally dispersed uniformly throughout the stream depth but large grains are more concentrated near the bottom.

There is a correlation between the suspended load and streamflow discharge for most streams. During stormflow events, the rising limb of the hydrograph is associated with higher rates of sediment transport and degradation (Fig. 9.2). As the peak passes and the rate of streamflow discharge drops, the amount of sediment in suspension also diminishes rapidly and aggradation occurs. If sufficient measurements of streamflow discharge and

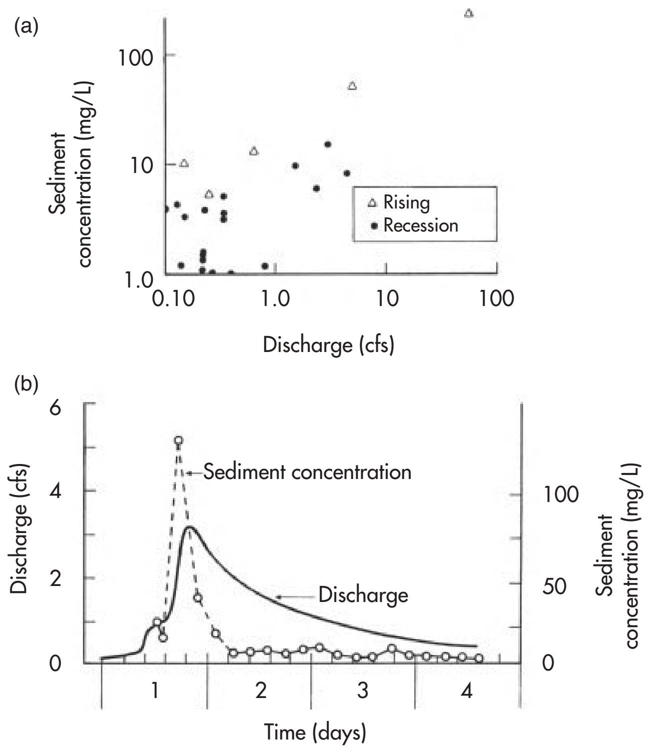


FIGURE 9.2. Example of the relationship between streamflow discharge and suspended sediment for a small stream in northeastern Minnesota. a: the relationship for several storm events; b: a single storm event (included in A)

sedimentation are available, a relationship can be developed for use as a *sediment-rating curve*. This relationship will most often be a power function form such as:

$$SS = kq^m \tag{9.9}$$

where SS is the suspended sediment load (mg/L); q is the rate of streamflow discharge (m^3/s); and k and m are the constants for a particular stream.

The variability in the sediment-rating curve for a stream is related to channel stability and sources of sediment. Streams draining undisturbed forested watersheds are characteristically stable with low levels of suspended sediment. However, both suspended sediment and bed load are discharged in streams draining forested watersheds depending on the stream type (see Chapter 10). The sediment-rating curve for a channel reach will change as a result of changes in the magnitude of sediment contributed from a watershed, the streamflow discharge regime, or the stream-channel morphology. For example, a wildfire can temporarily cause increases in the generation of suspended sediment downstream. The increase in the amount of sediment and streamflow associated with the wildfire can shift the relationship depicted by a sediment-rating curve (Fig. 9.2). Likewise, floods can change the relationship by bank overflow and the cutting of new channel segments that provide new sources of sediment. Newly cut stream channels can have sediment rating curves that are different from those of a well-armored stable stream. Suspended sediment-streamflow discharge relationships following such disturbances will eventually adjust back to the original sediment-rating curve–often requiring several years.

A sediment-rating curve such as obtained by Equation 9.8 can be used for estimating the effects of watershed management activities and land use on suspended sediment. To use such an equation, stable relationships must be developed from field data and any changes in the relationship caused by natural phenomena taken into account. Any significant shift in the relationship following some action such as timber harvesting or the conversion of vegetative cover on a watershed could then be quantified. A major difficulty, however, is separating changes in sediment rating curves caused by natural phenomena from those caused by people's activities. Obtaining a "representative" sample of suspended sediment is also difficult because concentrations can vary with time and within a cross section of a stream.

The usefulness of sediment rating curves is often improved by separating the data into streamflow generation mechanisms such as rainfall events or snowmelt runoff, rising and falling stages of the hydrograph, or combinations thereof (Box 9.1). The measurement and interpretation of suspended sediment data become important when determining water-quality compliance. If suspended sediment is above a water-quality standard of a specified state in the United States, a TMDL analysis will be required to bring the water body back into compliance. This topic is discussed in more detail in Chapter 11.

Bed Load. Bed load particles can be transported singly or in aggregates and can be entrained if the vertical velocity of eddies creates sufficient suction to lift the particle from the bottom. These particles can also be started in motion if the force exerted by the water is greater on the top of the grain than on the lower part. Particles can move by saltation if the hydrodynamic lift exceeds the weight of the particle (Fig. 9.1). The particles will be redeposited downstream if not re-entrained. Large and small particles can roll or slide along the stream bottom. Rounded particles are more easily moved. The largest particles are generally moved in the steeply sloping channels of headwater streams.

The largest grain size that a stream can move as bed load determines *stream competence*. The competency of a stream varies throughout its length and with time at any point along its length. Stream competence is increased during high peak discharges and flood events. The force required to entrain a given grain size is called the *critical tractive force*. *Erosion velocity* is the velocity at which entrainment of bed particles occurs. DuBoy's equation is generally used to calculate the tractive force for low velocities and small grains as a function of stream depth and gradient as follows:

$$T_f = W_w D S \qquad \textbf{(9.10)}$$

where T_f is the tractive force; W_w is the specific weight of water; D is the depth of water; and S is the stream gradient.

For high velocities and large particles, stream velocity is more important than depth and slope. This situation has given rise to the sixth-power law that is:

$$\text{competence} = CV^6 \qquad \textbf{(9.11)}$$

where C is a constant.

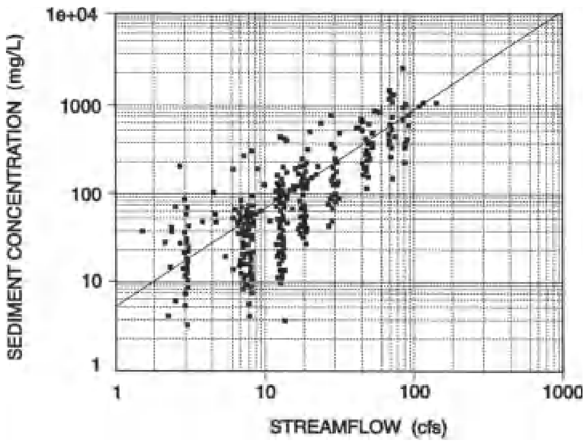
Doubling the stream velocity means that particles 64 times larger can be moved. However, the exponent is only approximate and varies with other conditions of flow.

Box 9.1

Sediment rating curves for watersheds of Arizona

Sediment rating curves were developed to analyze the effects of vegetative management practices on suspended-sediment discharge from watersheds in ponderosa pine forest and pinyon-juniper woodlands of north-central Arizona. The sediment rating curves were derived from measurements of suspended-sediment concentration and streamflow discharge on the Beaver Creek watersheds (Lopes et al., 2001) and are expressed in the form of power equations (Equation 9.9). One example of these curves is presented below. Disturbances from vegetation management practices generally increased suspended-sediment transport above those of control watersheds. Completely cleared and strip-cut ponderosa pine watersheds produced higher sediment concentrations than did a control watershed. Likewise, cabled and herbicide-treated pinyon-juniper watersheds yielded higher sediment-laden streamflows than did a control.

Sediment transport regimes were also related to streamflow-generation mechanisms and hydrograph stages. Although about 85% of the data analyzed represented snowmelt-runoff events in both vegetative types, derivation of sediment rating curves based on streamflow generation mechanisms (snowmelt-runoff events, high-intensity and short-duration rainfall events, and low-intensity and relatively long rainfall events) improved the sensitivity of the analysis. Sediment data collected during rising and falling stages of a hydrograph varied between the two vegetative types. Sediment concentrations were generally higher in the rising stage than in the falling stage for ponderosa pine watersheds. However, there was no clear evidence of higher sediment concentrations in the rising stage of the hydrograph as compared to the falling stage in the pinyon-juniper watersheds.



Relationship between streamflow discharge and suspended sediment from a clearcut ponderosa pine watershed in north-central Arizona (from Lopes and Ffolliott, 1993, © *Water Resources Bulletin*, by permission).

Stream power, the rate of doing work, expresses the ability of a stream to transport bed load particles. It is the product of streamflow discharge, water surface slope, and the specific weight of water. Relationships similar to a sediment rating curve (see above) can be developed between unit stream power and unit bed load transport rate for a stream.

Another important concept in sediment transport is *stream capacity*, which is the maximum amount of sediment of a given size and smaller that a stream can carry as bed load. Increased channel gradient and discharge rate generally result in increased stream capacity. If small particles are added to predominantly coarse streambed material, the stream capacity for both large and small particles is increased. But, if large particles are added to small-size grain material, the stream capacity is reduced. Small particles increase the density of the suspension and, therefore, the carrying capacity. The carrying capacity also decreases with increasing grain size.

All of the variables affecting stream capacity are interrelated and vary with channel geometry. Streams that carry large bed loads, for example, those in dryland regions where sediment sources are extensive, have shallow, rectangular, or trapezoidal cross sections because there usually is a steep velocity gradient near the streambed in the cross sections (Morisawa, 1968). In contrast, the typically parabolic cross sections of channels in humid regions where sediment loads are comparatively small usually do not have steep velocity gradients near the streambed.

Degradation and Aggradation

The amount of sediment carried by a stream depends largely on the interrelationships among the supply of material to the channel, characteristics of the channel, the physical characteristics of the sediment, and the rate and amount of streamflow discharge. The supply of sediment material and streamflow depend on the climate, topography, geology, soils, vegetation, and land-use practices on the watershed. Channel characteristics of importance are the morphological stage of the channel, roughness of the channel bed, bed material, and steepness of the channel slope. Soils and geological materials of the watershed and stream channels and the state of their weathering largely determine the physical characteristics of the sediment particles.

The interrelationships among these factors determine the type and amount of sediment and the amount of energy available for the stream to entrain and transport the particles. When stream energy exceeds the sediment supply, channel *degradation* occurs (Fig. 9.3). Localized removal of channel bed material by flowing water is called *channel scour*. On the other hand, *aggradation* occurs within the channel when sediment supply exceeds stream energy. For a particular stream and flow condition, a relationship between transport capability or capacity and supply can be developed (Fig. 9.4). The wash load illustrated in Figure 9.4 consists of silts and clays 0.0625 mm or smaller. Sediment supply generally limits total sediment transport for smaller particles but as the material gets larger, total sediment transport is more likely to be limited by transport capability.

The processes of degradation and aggradation are important when considering stream dynamics because these are the main mechanisms for sediment storage and release, respectively, in a stream channel (Fig. 9.3). When aggradation occurs, excess material is deposited and eventually a new slope is established that equals the upstream slope. The equilibrium of the new slope established by aggradation can carry the incoming sediment but the downstream slope has not adjusted and deposition occurs. This obstructs the flow

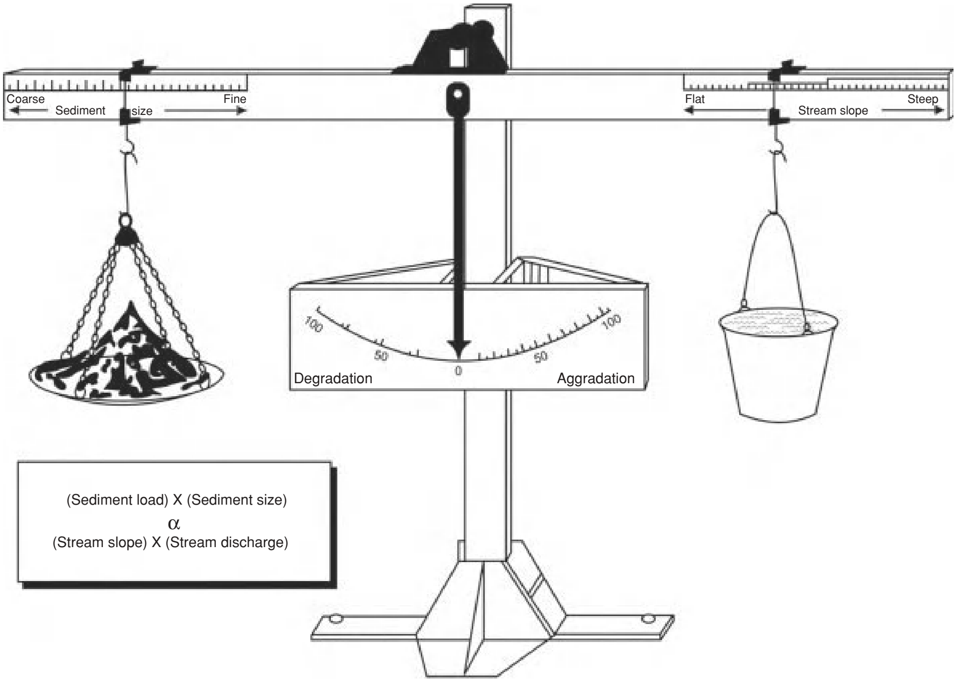


FIGURE 9.3. Relationships involved in maintaining a stable channel balance (from Lane, 1955, based on Rosgen, 1980)

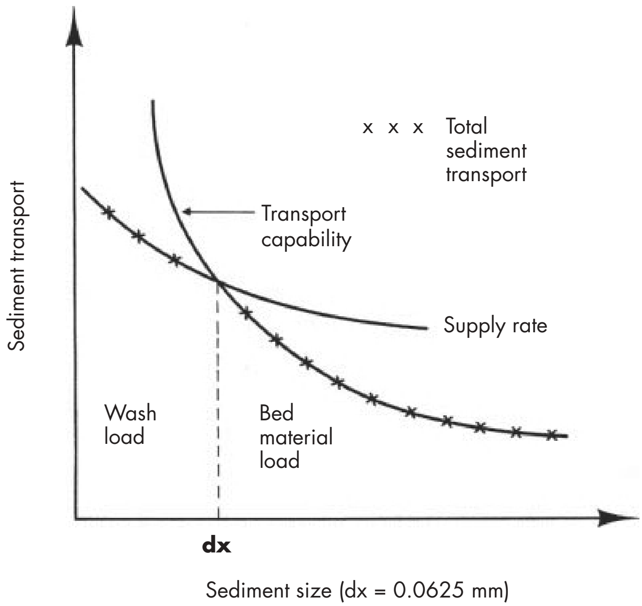


FIGURE 9.4. Rate of sedimentation as affected by transport capability and supply rate for different-sized particles for a particular stream and flow condition (from Shen and Li, 1976, as presented in Rosgen, 1980)

and deposition also occurs above the reach. As a result, deposition occurs above and below the reach to raise the bed parallel to the new equilibrium slope. The rate of this movement decreases with time or with downstream advance and so does the rate of aggradation. The channels upstream and downstream from the aggrading front behave like two different reaches with different flows until aggradation ceases and the status of dynamic equilibrium (see later) is restored.

Degradation involves sediment removal. Relatively speaking, natural degradation is usually a relatively slow process. However, rapid degradation can occur when the dynamic equilibrium of a stream system is disturbed by a change in land use. This initially rapid degradation following disturbance diminishes slowly through time. The profile of a degrading stream system is concave and the channel cross sections tend to be V-shaped down cut. Soil particles are picked up from the bed until the load limits are reached during the degradation process. The formation of V-shaped cross sections occurs because of the variations in flow and flow resistance across the channel. The carrying capacity for sediment is normally lower near banks than in the center of the stream channel because of bank roughness. Therefore, more material is picked up in the center of the stream, causing the V-shaped cross section.

Aggradation generally creates a convex-longitudinal profile in a stream system where coarse material is deposited first, while the finer material moves farther downstream. Particle size decreases downstream as a consequence. When aggradation occurs, the streambed rises slowly, usually mid channel, and there is a tendency for the water to flow over the banks due to a loss of channel capacity. This process can lead to natural levee formation. The continual deposition and aggradation can ultimately lead to a braided river.

Degradation and aggradation processes also lead to the shifts in channel morphology to be discussed in the following chapter.

Dynamic Equilibrium

Stream channels are in a constant state of change because of the processes of degradation and aggradation as suggested in Figure 9.3. When a channel system is in *dynamic equilibrium*, it is sufficiently stable so that compensating changes can occur without significantly altering this equilibrium (Heede, 1980; Schumm, 2005). This resilience or resistance to rapid change is caused by internal adjustments to change in flow or sediment movement that are made by several factors including vegetation, channel depth, and stream morphology operating simultaneously in the system. Dynamic equilibrium can also be explained in terms of hydraulic geometry.

Hydraulic geometry relationships (Singh et al., 2003) suggest that the spatial variation in stream power is accomplished by equal spatial adjustments between flow depth and flow width. As flow depth increases within limits, channel width must also increase assuming friction or roughness is fixed to allow equilibrium conditions to occur in the channel. In theory, the concept of dynamic or quasi-dynamic equilibrium under steady-state conditions tends to maximize entropy (Jaynes, 1957). Maximum entropy of a system occurs when the change in stream power is distributed among the changes in flow depth, channel width, flow velocity, slope, and friction factors or resistant materials as determined from particle size (Langbein, 1964; Yang et al., 1981). These processes can be illustrated by the principle of tractive force theory expressed in Lane's (1955) continuity equation, where sediment size and sediment load are proportionally balanced by streamflow discharge and channel slope (see Fig. 9.3).

Streams in dynamic equilibrium do not have headcuts. Watercourses that begin high on the watershed form a smooth transition between the unchannelized area and the defined channels. Bed scarps are not developed and the longitudinal profile is concave. In general, flow discharge, depth, and width increase downstream, and gradient and sediment particle size decrease, if watershed conditions are relatively constant over long stream reaches. Sediment production is negligible as a consequence.

Streams that are not in equilibrium are indicated by channel headcuts, underdeveloped drainage nets such as those having channelized watercourses on only one-half or less of the watershed area, frequent bed scarps, and the absence of a concave-longitudinal profile where watershed conditions are relatively constant. Channel headcuts are sources of local erosion and indicate that the stream length and gradients have not allowed an equilibrium condition to develop. Bed scarps develop at knickpoints and indicate marked changes in longitudinal gradients. These scarps proceed upstream until a smooth transition between upstream and downstream gradient is attained.

Measurement of Sediment

Measurements of suspended sediment and bed load are obtained by a number of methods. Some of these methods are also used to collect samples for other water-quality analyses.

Suspended Sediment

Obtaining grab samples is a common procedure of measuring suspended sediment in small streams. However, this method is not always reliable because of the variability in suspended sediment concentrations. Single-stage samplers consisting of a container with an inflow and outflow tube at the top are used on small fast-rising streams. A single-stage sampler begins its intake when the water level exceeds the height of the lower inflow tube and continues until the container is full. The use of such data is limited because only the rising stage of the hydrograph is sampled.

Depth-integrating samplers minimize the sampling bias encountered with single-stage samplers. A depth-integrating sampler such as the DH-48 or DH-49 has a container that allows water to enter the sampler as it is lowered and raised at a constant rate. As a result, a relatively uniform sample of a vertical section of a stream is obtained. Measurements are obtained by standing in smaller streams or from a cable car suspended above larger streams and rivers. Depending on the size of the stream, a number of these samples are taken at selected intervals across the channel. Each measurement should also be accompanied by a measurement of streamflow discharge through the channel cross section.

A variety of pumping samplers that automatically collect a sample of water at a specified time interval from a point in a stream are also available to analyze suspended sediment concentrations. The components of these samplers are the intakes, a pump, a splitter that draws off a specified volume of water, a circular table holding the containers in which the samples are pumped, and a water supply for priming the pump and flushing sediment out of the intake before each sample is taken. There is also a clock and control box with a timer that initiates the sequence of operations in proper order. With the exception of the intakes that are positioned at a point in the stream channel, these components are housed in a shelter.

It is preferable that the pumping sampler be located adjacent to a continuous water-level recorder to obtain a measure of the water level of the stream when each sample is collected. The advantage of a pumping sampler is that samples of stream water are collected over a specified time period when a person is unavailable to collect samples.

Regardless of the methods used after a suspended-sediment sample is collected, the liquid portion is removed by evaporating, filtering, or centrifuging and the amount of sediment is weighed. The dry weight of suspended sediment is generally expressed as a concentration in (mm/L) or parts per million (ppm) of water. Measurements of suspended sediment and bed load are usually made separately because of differences in the sizes of the particles and in the distribution of particles in a stream.

Bed Load

Bed load is more difficult to measure than suspended load. No device for measuring bed load is reliable, economical, and easy to use. While many bed-load samplers exist, the hand-held Helley-Smith bed-load sampler is one of the most widely used on small streams in the western USA (Leopold and Emmett, 1976). Estimates of bed load can also be obtained by measuring the amount of material deposited in sediment traps, settling basins such as found upstream of weirs, upstream of porous sediment-collecting dams, or reservoirs. These volumetric measurements can then be partitioned into sands, gravels, and cobbles to determine the contributions by particle size. The USGS has been experimenting with ADCP data to better understand bed load movement.

Total Sediment

Measurements of total sediment can be obtained in an installation consisting of a low dam and basin to trap the coarse sediments in a bed load and a series of splitters that collect a known portion of the suspended sediment passing over a spillway in the dam (Brown et al., 1970). Low volumes of streamflow that do not spill over the dam deposit their sediment in the basin. However, smaller sediment particles might pass over the spillway during intermediate streamflow events with increasing larger particles passing through the spillway in larger flows and, as a consequence, these sediment particles are not sampled. It is necessary, therefore, that the basin be designed so that it does not completely fill with sediment for the anticipated future streamflows from the watershed. Following cessation of streamflow, the basin is drained of the remaining water and the bed-load sediments collected for measurement and analysis. Use of this procedure to measure bed load is best suited to intermittent or ephemeral streamflow regimes because of this need to drain the basin.

Continuous measurements of suspended sediment are obtained with the use of a splitter or series of splitters. As water flows through the spillway of the dam, or through a weir, a small fraction is separated from the flow by a splitter (Fig. 9.5). Where large streamflow discharges occur, the flow can be further sampled with a second and even a third splitter. An intermediate storage tank is generally placed between the second and third splitter with the third split from this tank transporting water into a final storage tank. A screen is placed at the outlet of the intermediate tank to prevent trash from reaching the third splitter. The water collected in the final storage tank is then sampled for measurement and analysis.

FIGURE 9.5. A sediment splitter diverts progressively smaller fractions of streamflow to a collection tank to sample suspended sediment flowing through a rectangular weir in north-central Arizona. By also collecting bedload materials that settle in the catchment basin upstream of the weir total sediment yields can be measured from a watershed (For a color version of this photo, see the color plate section)

The fraction of the total suspended sediment sampled is determined through a calibration of the splitters. If the proportion split by the first splitter is 1:600, the proportion of the split by the second splitter is 1:10, and the proportion of the split by the third splitter is also 1:10, the proportion of the water flowing through the spillway and entering the final storage tank is 1:60,000. In this example, the suspended sediment sampled in the tank would be multiplied by 60,000 to obtain an estimate of the total suspended sediment in the streamflow for the period of data collection. Necessary assumptions in the analysis of the sample are that the respective proportions of the splits are constant for all streamflow discharges and that the splitters divert an unbiased sample of the average suspended sediment concentrations of the water flowing over the spillway for all concentrations encountered.

The weight of the bed load collected in the basin is added to the weight of the suspended-sediment load that has passed through the spillway to compute total sediment outflow from the watershed. If the bed-load deposition is small, it can be weighed directly and adjustments made for its moisture content. The dry weight of large depositions of bed-load materials is determined by volumetric surveys and measurement of the weight per unit volume. The suspended sediment computations involve assumptions that the weight of the sediment in the intermediate and final storage tanks can be divided by the volumetric split-proportion entering the respective tanks and that the calculated total loads for each tank are additive.

TABLE 9.2. Average annual sediment yield from selected river basins that are among the 21 largest sediment yielding rivers in the world

River, country	Drainage area ($\times$ 10^6 km^2)	Average annual streamflow discharge ($\times$ 10^9 m^3/year)	Average annual sediment yield (10^6 t/yr)[a]
Ganges/Brahmaputra, India	1.48	971	1670
Yellow (Huangho), China	0.77	49	1080
Amazon, Brazil	6.15	6300	900
Mississippi, USA	3.27	580	210
Mekong, Vietnam	0.79	470	160
Nile, Egypt	2.96	30	110
La Plata, Argentina	2.83	470	92
Danube, Romania	0.81	206	67
Yukon, USA	0.84	195	60

Source: From Milliman and Meade (1983), by permission.
[a] Metric tons per year.

SEDIMENT YIELD

Sediment yield is the total sediment outflow from a watershed for a specific period of time and at a defined point in a stream channel. Sediment yield is normally determined by sediment sampling and relating the measurements to streamflow discharge or conducting sediment deposit-surveys in reservoirs. Sediment fingerprinting has become a more commonly used tool to estimate and apportion sediment sources within a watershed (Engstrom et al., 2009). Estimated sediment yields from major river basins of the world are presented in Table 9.2 .

Streams discharging large quantities of sediment are those that drain areas undergoing active geologic erosion or intensive land use that can cause large erosion rates on watersheds and in channels. Sediment yields in relation to streamflow discharge are generally higher in dryland regions than more humid regions because of the lower vegetation densities of protective vegetation and the consequent higher rates of soil erosion (Box 9.2).

Undisturbed forested watersheds generally yield the lowest amount of sediment of any vegetative cover or land-use condition. Examples of annual sediment yields from small forested watersheds compared to larger watersheds with mixed land uses in the United States are presented in Table 9.3 .

Sediment Budgets

As suggested earlier, the transport, or routing, of sediment from source areas where active erosion takes place to downstream channels involves many complex processes. However, a *sediment budget* is an accounting of sediment input, output, and change in storage for a particular stream system or channel reach. Therefore, a sediment budget is a simplification of the processes that affect sediment transport and includes consideration of:

- the sources of sediment (surface erosion, gully erosion, soil mass movement, bank erosion);

Box 9.2

A comparison of sediment yields in two large river basins

The Mississippi River basin is 4 times larger than that of the Yellow River in China and the annual streamflow discharge is 12 times greater. However, the ratio of the sediment load to streamflow discharge for the Yellow River is 22.04 compared to the 0.36 ratio for the Mississippi River (see Table 9.2). The Mississippi River traverses a mostly humid region–much of the basin is vegetated naturally or agricultural crops and soil conservation is practiced over extensive areas. In contrast, the Yellow River drains a semi-arid region of north-central China, an area of deep loess, and is highly susceptible to geologic erosion. Furthermore, much of the basin has been denuded by centuries of "primitive" agriculture.

- the rate of episodic movement of sediment from one temporary storage area to another;
- the amount of sediment and its time in residence in each storage site;
- the linkages among the processes of transfer and storage sites; and
- the changes in sediment material as it move through the system.

A sediment budget is a quantitative statement of rates of production, transport, and discharge of sediment. To account properly for the spatial and temporal variations of transport and storage requires sophisticated models. While a discussion of such models is beyond the scope of this chapter, references or URLs for details on these models are listed in the bibliography and webliography, respectively, at the end of this chapter.

Table 9.3. **Sediment yield from small forested watersheds and larger watersheds of mixed land use in the United States**

		Sediment yield (t/ha/year)[a]	
Region	Number of watersheds	Mean	Range
East			
Forested	65	0.17	0.02–2.44
Mixed use	226	0.35	0.02–4.42
West			
Forested	80	0.16	0.02–1.17
Mixed use	312	0.42	0.02–13.38
Pacific Coast			
Forested	26	3.93	0.04–43.56
Mixed use	103	10.37	0.13–111.86

Source: From Patric et al. (1984).
[a] Metric tons per year.

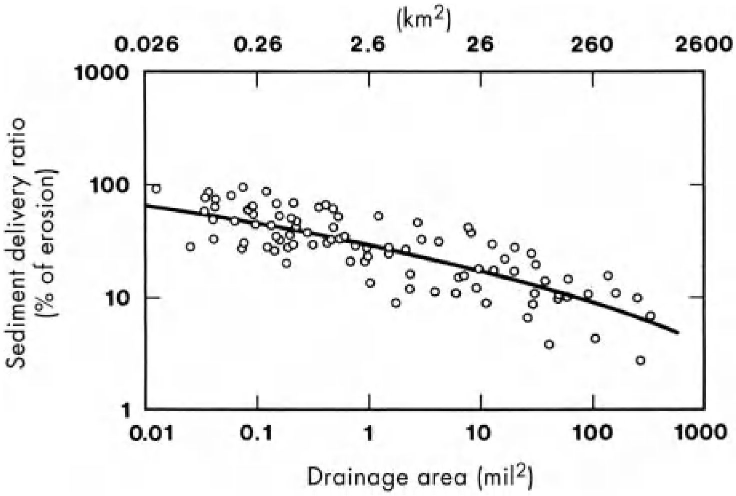


FIGURE 9.6. Sediment delivery ratio determined from watershed size (from Roehl, 1962)

Sediment Delivery Ratio

A commonly used method of relating soil erosion rates to sediment transport is the calculation of the *sediment delivery ratio* (D_r), defined as:

$$D_r = \frac{Y_s}{T_e} \tag{9.12}$$

where Y_s is the sediment yield at a point (weight/area/year); and T_e is the total erosion from the watershed above the point at which sediment yield is measured (weight/area/year).

The sediment delivery ratio is affected by the texture of eroded soil material, land-use conditions, climate, local stream environment, and general physiographic position. Generally, as the size of the drainage area increases, the sediment delivery ratio decreases (Fig. 9.6). Such relationships should only be used to provide approximations because erosion and sediment concentrations can vary greatly for a watershed.

Erosion and sediment data must be collected or otherwise be available before sediment delivery ratios or more detailed sediment-routing models can be developed. Often, these data are not available for upland watersheds.

Cumulative Watershed Effects on Sediment Yield

Sediment yield can be altered by any changes on a watershed that influence sediment deposition to a channel, streamflow rates, or both. Likewise, any changes to the channel itself that affect stream slope, channel roughness, and channel morphology can alter aggradation–degradation processes and sediment delivery. Among the array of people's activities on watershed landscapes that can affect these relationships are altering vegetative cover, agricultural crop cultivation, urbanization, road construction and maintenance, loss of riparian communities, wetland drainage, ditching, and stream channelization (Dissmeyer, 2000). However, these changes often occur piecemeal and are implemented incrementally over time. Nevertheless, the cumulative effect of the changes can impact the dynamic equilibrium of streams (Box 9.3).

Box 9.3

Cumulative watershed effects on stream sediment: two examples in the United States

Minnesota River Basin

The 43 264 km^2 Minnesota River Basin is the largest tributary to the Mississippi River in Minnesota. The 558 km river flows through some of the state's richest agricultural land. Like many basins in the midwestern USA, the basin has undergone major changes since the arrival of European settlers. Conversion from prairie and savannah vegetation to agricultural cropping, drainage of wetlands, urbanization, construction of roads, construction of levees, river channelization, removal of native riparian forests, and development on floodplains have changed the landscape and hydrologic characteristics, sediment loads and associated turbidity of the basin (Lenhart et al., 2011). Over time, the rates of erosion and the magnitude of streamflow volumes and peaks have increased, resulting in greater sediment levels in the river. Thirteen years of data at Mankato indicated a median suspended sediment concentration of 92 mg/L (Magner et al., 1993) and an annual sediment load of 26 metric tons/km^2 (MPCA, 1994). These sediment loads have occurred in spite of improved agricultural cultivation and conservation practices that have reduced surface soil erosion in much of the basin. Higher sediment loads in the 1980s compared to the 1970s can partly be explained by higher river flows (Klang et al., 1996). It is suspected that the extensive loss of wetlands and the associated ditching and drain tiling in the basin have cumulatively increased streamflows (see Chapter 14).

Nemadji River

The Nemadji River which drains an area of about 1120 km^2 that straddles the Minnesota-Wisconsin border south of Duluth, Minnesota, has the highest suspended sediment load per square kilometer of all monitored rivers in either Minnesota or Wisconsin (NRCS, 1998). More than 118,000 metric tons of suspended sediment and 4000 metric tons of bedload are discharged into Lake Superior annually. Hillslopes are unstable and soil mass movement is common in this area that is dominated by clay deposits and clayey-glacial till. Since the middle 1880s, sediment deposition from the Nemadji River into Lake Superior has increased. Changes in watershed conditions that possibly explain the increased sedimentation include (1) intensive timber harvesting of mostly white (*Pinus strobus*) and red pine (*P. resinosa*) forests in the late 1800s and the associated logging activities; (2) channel cleaning, straightening, and dam building that occurred in conjunction with logging;

(3) replacement of the pine forests with pioneer aspen (*Populus* spp.) forests and agricultural cropping; (4) road construction and maintenance; and (5) degradation of riparian areas.

The cumulative effects of these activities include increased streamflow volumes and peaks, wetter soils and accelerated rates of soil slumping along hillsides adjacent to channels, unstable stream channels, and consequently, increased channel erosion (Riedel, 2000; Riedel et al., 2005), and increased sediment transport into Lake Superior. Reversing these cumulative effects will require a comprehensive restoration program that includes efforts to (1) restore pine forest cover, (2) improve road maintenance and drainage, (3) manage riparian areas, and (4) restore stream channels (as discussed in Chapter 10).

Many studies describe the effects of various land-use activities on sediment yields without identifying the sediment source. For example, sediment yields attributed to timber-harvesting activities and roads are widely documented (Reid, 1993; Megahan and Hornbeck, 2000, and others). These studies generally show up to a 50-fold increase in sediment yields with most of the increase associated with improperly aligned roads. Increases in sediment input can be much larger on sites where landslides are common. Sediment yields decrease relatively rapidly once road use is discontinued and logged areas regenerate.

Studies that considered only the effects of roads on sediment delivery to stream systems in the Oregon Coast Range, USA, indicated that road density alone could not explain the rates of sediment yield from surface erosion (Luce et al., 2001). These studies suggested several direct and indirect effects that roads can have on sediment delivery including:

- road slope, length, and condition of the surface as affected by traffic and maintenance;
- distance from the stream channel and the road drainage system (culverts, ditches, etc.) that affects the volume and rates of sediment and water entering stream channels;
- gullies and mass wasting initiated by road construction and drainage on hillslopes;
- failures of culverts due to blockage or poor design;
- effects of roads on peak flows caused, for example, by concentrating surface runoff from the road surface or intercepting subsurface flow; and
- roads located in riparian corridors or on floodplains and cut and fill activities and use of riprap that changes the channel form.

Urbanization increases sediment yield in many instances (see Chapter 14). For example, Walling and Gregory (1970) reported up to 100-fold increases in suspended sediment concentrations below building construction sites. In a more specific example, Wolman and Schick (1967) estimated increases in sediment yields of up to 1800 t per 100 inhabitants because of construction activities in metropolitan Washington, DC.

While sediment yields are often an indicator of land-use impact, they are also difficult to interpret in terms of causal effects without information on how the sediment was produced.

Van Sickle (1981) summarized sediment yields from a large number of Oregon watersheds and showed that year-to-year variations were large. To arrive at an accurate long-term average value, therefore, it is necessary to measure sediment yields over a long-time period. Roels (1985) have stated that many of the results obtained from plot-sized experiments cannot be used to estimate soil erosion rates and sediment yields on a watershed basis because of the inherent limitations in the study designs. It is important, therefore, that all sediment measurement and monitoring programs be carried out within the framework of statistically valid sampling schemes.

The diversity of land-use changes occurring spatially and temporally on a watershed can add up to significant changes in sediment yield from the watershed. For example, reductions in or elimination of riparian vegetation and streambank alterations that increase streambank erosion can increase sediment yield. Increases in streamflow resulting from reduced time of concentration of runoff can lead to changes in stream-channel dimensions to accommodate the modified flow. Channel erosion, primarily lateral extension, which increases channel width, adds to sediment supply. Reductions in flow can also lead to channel adjustments and alterations of sediment yield.

Summary and Learning Points

Surface soil erosion, gully erosion, soil mass movement, streambank erosion, and channel scour combine to produce the sediment supplied to stream channels. Streams transport this sediment as suspended sediment and bed load. The channel processes involved in sediment transport and deposition are dynamic and complex. Important fluvial processes and hydraulic variables are involved in the aggradation and degradation of stream channels during the storage and subsequent release of sediment in stream channels. The general purpose of this chapter was to introduce the subject and provide sufficient information so that you should be able to:

1. Explain the fundamental energy relationships of streamflow in a channel that affect streamflow velocity, stream-channel erosion, and sediment transport.
2. Describe the different sediment sources and the types of sediment transport.
3. Explain the relationships between stream capacity and sedimentation.
4. Explain the conditions under which aggradation and degradation occur in a stream channel.
5. Discuss the relationships that are generally found when the hydrograph characteristics of stormflow events are compared to the corresponding suspended-sediment loads.
6. Explain the relationships between upland soil erosion and sediment delivery to streams.
7. Identify the factors affecting a sediment delivery ratio.
8. Explain the differences between laminar and turbulent flow, and between subcritical and supercritical flow.
9. Discuss why dynamic equilibrium is a useful concept in describing stream systems and their stages of development.
10. Provide examples of how cumulative watershed effects can alter sediment transport and delivery.

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CHAPTER 10 Fluvial Processes and Implications for Stream Management

INTRODUCTION

Over geologic time, there is an ongoing struggle on the landscape between the uplifting of landmasses and the weathering processes of erosion and sediment transport that work to lower landmasses to sea level. Rock and soil have varying degrees of resistance to erosion depending on their mineralogy and the forces working to change the landscape. This section of the chapter will focus on landforms shaped by water that is the purview of *fluvial geomorphology*. The dynamic equilibrium between water and sediment described in the previous chapter determines the form of a valley landscape and delineates watershed boundaries and its embedded stream channels.

Flowing water alters the topography and forms stream channels that are the conduits for moving water and sediment. In recent decades, hydrologists and watershed managers have increasingly recognized the importance of fluvial geomorphology. This perspective helps them to better understand how watershed-management practices and land-use activities affect the flow of water and sediment. Through field observations of land and channel forms, it is possible to link geomorphic features with hydrologic processes and, furthermore, link people-induced changes on a watershed with changes in the stream channel.

The intent of this chapter is to link hydraulic principles and fundamental energy relationships with valley type and stream channel form, function, and adjustment with time. Our focus is on *how* inherent valley features, management practices, and land-use activities influence stream channel morphology and the resulting flow of water and sediment. We provide an overview of valley- and stream-classification systems and how they are related to land-use activities and management practices and their effects on hydrologic function in different geomorphic settings. The purpose is not to promote classification systems themselves, but rather to show how these classification systems offer implications for planning

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and managing streams in a watershed context. An understanding of fluvial processes and classification systems can help managers determine how climatic and watershed conditions affect stream channels and, for example, assist them in making better decisions about spending limited funds for channel improvements and restoration.

FLUVIAL GEOMORPHOLOGY

A stream channel is formed by the energy of flowing water and the factors that affect that flow including topographic relief, geologic strata, parent material and soils, vegetation, and characteristics of the riparian corridor. Water initially flows over the land surface or through the micro- and macropore systems of soil and rock. These flows begin to cut a channel at the uppermost parts of a watershed where flow becomes concentrated and has sufficient kinetic energy to pick up and transport sediment. The channel can be a small rivulet at first that then combines with other rivulets to gain greater velocity. The channel can be fed by subsurface flow from the hillslope and channel bank, overland flows of water, and groundwater as it continues downstream as described in Chapters 5 and 7.

A stream or river can be viewed as having an upper-erosion zone, a transition zone, and a deposition zone (Schumm, 1977) as shown earlier in Figure 1.2. The erosion zone is the upper reach of a river basin and generally has the steepest hydraulic gradient unless the basin was formed by glacial scour. With the coincident increase in mass of streamflow, the channel deepens, widens, and begins its course to the ocean, as per the energy relationships of water and sediment flow discussed in previous chapters. The main stem of the river is formed and is considered a transition zone between the upper-erosion zone and the lower-deposition zone at the river estuary. There is little gain or loss of sediment in the transition zone over the long term. The deposition zone has a low hydraulic gradient as evidenced by the vast deltas formed by major rivers such as the Mississippi, Congo, and Nile.

As the magnitude of flow increases during rainfall- or snowmelt-runoff events, there is a corresponding increase in energy for picking up and transporting sediment. The increased energy of large flows is allocated into work on the channel itself and in transporting sediment in the channel. Over many years of episodic stormflow events followed by longer periods of baseflow, channels in the reaches of a river basin take on their form. The channels will adjust their form depending on their position in the basin and the corresponding type of valley, the hydraulic gradient, the type of material present, and other factors to be discussed below. The adjustment of form is driven by process. The Channel Evolution Model (CEM) developed by Schumm (1977), Schumm et al. (1984), and Simon and Hupp (1986) provides a way to determine how most alluvial river valleys and channels can change in form over time.

The CEM illustrates how knickpoint migration, downcutting, and widening cause bank erosion and explains how deposition processes produce a new active floodplain along with terraces. Overbank deposition due to upland erosion (Trimble, 1983) can also produce incised channels that will need to enlarge through bank erosion before they reach a more stable form. The CEM is compared to other stream-classification systems later in this chapter.

Rivers, Floods, and Floodplains

Rivers and their tributaries form *channel banks* as they cut into the earth in their journey downstream. These banks form over time and become established at a height that confines the river to its channel for all but the larger streamflow events. When streamflow increases to the point where its elevation exceeds the elevation of the streambanks, part of the river leaves its channel and a flood occurs. The overbank flow spreads out onto the floodplain where its velocity is reduced because of the increased roughness encountered from riparian vegetation and debris deposits from previous floods. As a result, sediment and debris are deposited to form a relatively flat feature that is adjacent to the channel – this is the *floodplain*. The elevation at the top of the streambank where water begins to overflow onto the floodplain is the *bankfull stage*. Water in many river systems flows onto the floodplain every 1 to 2 years. Over time, the river in the upper erosion zone lowers its base elevation and new floodplains develop. *Terraces* are old floodplains that have been abandoned as the stream channel lowers its base elevation. These features are illustrated in Figure 10.1.

A river experiences floods of various magnitudes through time. Major flood events with deeper flow on the floodplain produce higher velocities, that can cause considerable soil erosion, and sediment transport. A new channel might be cut into the floodplain at some locations and, as the flood recedes, more sediment and debris are deposited. Streams meander as the hydraulic gradient is reduced; that is, they move laterally throughout the valley floor by eroding one bank and depositing sediment on the opposite side of the channel. Streams can migrate laterally within the floodplain as a result of both flood events and natural channel-forming processes. Over a long period of time, the stream channel will eventually have occupied all positions on the valley floor between the valley walls (Dunne and Leopold, 1978). As the channel meanders, it maintains its depth and width dimensions unless there is a substantial shift in either flow regimes or sediment deposition in the channel. Over time, new sections of channel are formed and other sections of the original channel abandoned – this is how most valleys are formed.

Bankfull Stage

The bankfull stage and its attendant discharge are morphological indices that can be related to the formation, maintenance, dimension, pattern, and profile of a channel under the current climatic regime (Rosgen, 1996). In the USA, streamflow discharges associated with the bankfull stage for most streams have a recurrence interval that averages 1.5 years (Simon et al., 2004). This discharge is the most effective at doing the work that maintains the channel's morphological characteristics. The reason for this effectiveness is illustrated in Figure 10.2. The largest streamflow discharges that have occurred historically in a channel possess the greatest power to form a channel and transport the most sediment; however, they occur infrequently. Nevertheless, the power of these infrequent streamflow events can move the large particles accumulated in the channel that would not be moved in more frequent events (Fig. 10.3).

Apart from these extreme events, the product of the transport rate and the frequency of occurrence at a maximum at some point between the most frequent discharge level of a stream and the highest levels of flow is the *effective discharge* (Wolman and Miller, 1960). The effective discharge is generally associated with that at the bankfull stage. This indicates that the flows at or near the annual maximum peak-flow discharge are the most important in "working" the channel and its sediments. As a result, the bankfull or channel-forming stage has become the key or "standard" for describing the morphological features of a stream channel such as its width, depth, and entrenchment. This standard became the basis for the

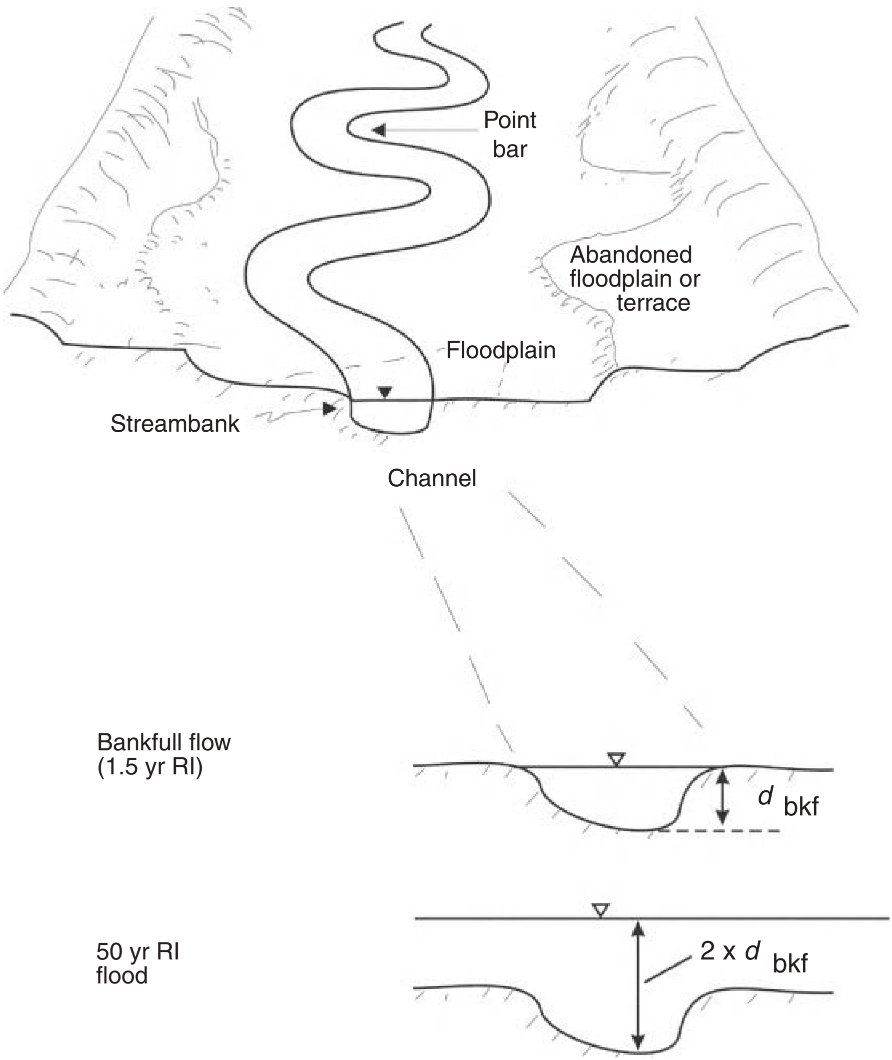

FIGURE 10.1. Landforms in a river valley and flow conditions in a channel for bankfull and the 50-year recurrence interval (RI) flood (modified from Fitzpatric et al. (1999) and Verry (2000))

morphological classification and stream-channel succession model proposed by Rosgen (1996, 2006) to be discussed later in this chapter.

One might be asked if the bankfull discharge for all streams coincides with the 1.5-year recurrence interval of the bankfull stage for most streams. Surprisingly, this seems to be consistent (Simon et al., 2004). However, Rosgen (1996) points out that stream channels with bed material consisting of large cobbles and boulders require streamflow discharges of a larger magnitude and with a longer return period than flows near the average annual event. Nevertheless, most stream channels seem to be formed and maintained by the more frequently occurring flood events rather than the extreme and infrequent floods.

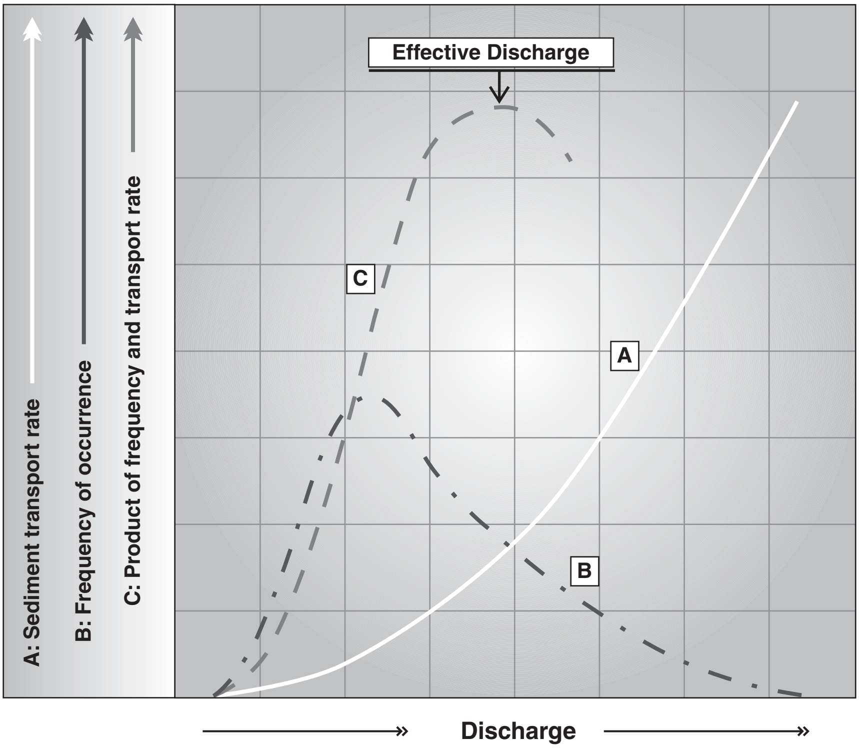

FIGURE 10.2. Effective discharge is determined as the maximum product of sediment transport rate and the frequency of occurrence of streamflow discharge (from Wolman and Miller (1960) as presented by Rosgen (1996). © Wildland Hydrology, with permission)

The Floodplain

The relationship between a river or stream channel and its floodplain is important. During flood events, the floodplain becomes an integral part of the river. As stated earlier, this is not a rare occurrence and a fact often ignored or not well understood by those who cultivate, urbanize, or otherwise develop enterprises on a floodplain. It is important to note here that there is a distinction between the floodplain and terraces that represent the old, abandoned floodplain (Fig. 10.1). While an active floodplain experiences frequent inundation, a higher-elevation terrace will be flooded less frequently if at all. As pointed out by Dunne and Leopold (1978), it is, therefore, important for developers to recognize the difference between the two.

When *overbank flow* occurs and moves onto floodplains unaltered by people's activities, the flow encounters a higher level of resistance or roughness that is due to natural riparian vegetation and debris from previous floods. The reduced velocity has hydrologic implications on a floodplain and downstream areas. In the immediate vicinity of the floodplain, a given streamflow discharge will have a higher stage as a result of the reduced velocity. This means that the area inundated by the flood is larger than it would otherwise be with a higher velocity. The accumulation of sediment and debris can further add to the

FIGURE 10.3. A large rainstorm event in August 18–19, 2007, produced nearly one-half of the annual precipitation within the Garvin Brook watershed, in southeastern Minnesota, USA. While this was an infrequent event (500-year), large boulders and cobbles in the stream channel and floodplain were redistributed by extreme boundary shear and a new channel was cut through a playground at Farmers County Park

height of the stage along with increases in roughness. For larger floods associated with 5-year, 10-year, 50-year, or longer recurrence intervals, large depositions of sediment can occur. The more extreme floods can scour a new channel and accelerate channel migration in the floodplain as mentioned previously.

By reducing the velocity of flow over the floodplain and providing storage, floodplains have a cumulative effect of reducing flood peaks downstream. Furthermore, there is evidence that the saturated soils and residual pools of water retained in the floodplain can reduce nitrogen levels that would otherwise be transported downstream (Cirmo and McDonnell, 1997). Uptake of nutrients by riparian vegetation and denitrification (see Chapter 11) can reduce nitrogen levels in the floodplain. Finally, the production of fisheries in the stream is strongly related to the amount of floodplain accessible by a water body (Junk et al., 1989).

VALLEY AND STREAM EVALUATION AND CLASSIFICATION

The form and functions of a valley and its stream channel affect the pattern of streamflow, the types of aquatic habitats present, and the quality of water in the stream. Likewise, people's activities on a watershed and within the riparian corridor and stream channels of the watershed can affect the stability, form and function of streams, quality of water, and aquatic ecosystem. Over time, there can be a stream reach or entire river system that becomes degraded or has undergone changes that indicate that a rehabilitative action is needed. Such changes can include excessive streambank erosion, sediment deposition in pools, increased suspended sediment concentrations, river widening, loss of fish habitats, and degraded aquatic ecosystems. A watershed manager needs to evaluate, classify, and set priorities for streams in response to such situations.

As stream channels adjust to either natural or people-induced perturbations, there is often interest in evaluating the status of streams and ultimately developing some type of remedy to rehabitate or improve form, habitat, and function of the stream. However, there are management limitations on some watersheds, for example, where there are naturally occurring high levels of sediment or degraded channels that are largely driven by natural phenomena. Magner and Brooks (2008) present the case of the Nemadji River in Minnesota, USA, where it is unlikely that a stringent cold-water standard for turbidity will ever be achieved due to groundwater entrainment of lacustrine silts and clays.

Deciding if management activities are warranted in a given stream or prioritizing stream sections or watershed areas in need of management actions have led to numerous approaches for evaluating and classifying stream channels in terms of their morphology, habitat, and related biological conditions. Because of the diverse nature of physical and biological relationships in a stream and the objectives of a particular watershed program, be they to meet water-quality standards, improve riparian conditions, restore channel conditions, or enhance fish habitats, there is no single classification system that applies for all interests. In contrast to the stream-morphology approach such as CEM discussed earlier some systems focus on the biological aspects of stream habitats and aquatic organisms (Hilsenhoff, 1987; Hawkins et al., 1993). In either case, there are criteria (Mosley, 1994) that should be noted in adopting a classification system including

- the purpose and applications of the classification must be specific and clear;
- objects such as stream channels that differ in kind should not “fit” into the same classification;
- the classification is not absolute and, therefore, can change when new information becomes available;
- the classification should be exhaustive and exclusive, and a stream channel should be assigned to only one class; and
- differentiating characteristics must be relevant to the purpose of the classification.

A detailed discussion of valley and stream evaluation and classification systems is not a purpose of this book. However, we will consider some of the classification and evaluation methods that are applied widely.

Valley Classification Systems

The climate, underlying geology, and land-use activities produce varying landscapes and landforms that control the development of streams within a valley. For example, the controlling factors of channels in dryland regions of the western USA result in discontinuous gullies on convex slopes in small drainage basins where tributaries contribute sediment to that accumulated in streams over time (Schumm, 2005). The lack of rainfall and vegetation and the presence of sand, gravel, and cobble in the streambed provide insight into the type of stream that has developed. In contrast, the more humid Lake States of the upper Midwest, USA, produce little sediment yield through flat terrain cohesive sediment (Magner et al., 2012) or stream systems underlain with resistant bedrock that carved relatively small channels into the resistant bedrock (Miller, 2011). The area of the resistant-bedrock channel is small compared to the cross-sectional area of a channel cut through soil because the shear forces acting on the streambed and banks require long time frames to enlarge the channel. Therefore, examining landscapes and landforms reveals much about how water and sediment will be stored and moved in a watershed. Describing valley features and providing some measure of classification is also helpful toward understanding stream channel dynamics.

Gibling's Classification System

Gibling (2006) developed a valley classification system that is based on bedrock, alluvial, eolian, marine, and glacial settings. The width (*W*) and thickness (*T*) of the valley conditions in more than 1500 fluvial bodies were noted by Gibling in developing this valley classification system. A range of *W*/*T* ratios from less than 1 to more than 1000 were found. These data were then classified into 12 valley-channel bodies. The *W*/*T* ratios were further classified based on

- mobile channel belts such as braided and low sinuosity systems and meandering river deposits;
- mixed channels and poorly channelized systems including large alluvial fan deposits;
- delta distributaries, for example, coastal wetlands cut into cohesive clay or organic material; and
- valley fills such as sediment fills over openings in bedrock or scoured alluvial, marine, or glacial deposits.

Gibling's Classification System describes a range of fluvial features found throughout the world. However, from a watershed management perspective, this system can be difficult to apply without expertise in geology. Therefore, we will also consider the Rosgen's classification approach that is widely used.

Rosgen's Classification System

Rosgen (1996, 2006) offers a valley- and stream-classification system that can be used in watershed management. Knowing the valley type limits the possible stream types, which is essential for assessing stream health and making recommendations for restoration. This system presents 11 different valley types based on their geology including bedrock, particle size, morphology, and slope features (Table 10.1). Valley type was incorporated into the classification system to provide a large-scale assessment (referred to as Level I) to limit the options for possible stream types and associated channel stability. The stability and activity of a stream at a reach are dependent on the location of a stream within a valley. Therefore, defining the valley type is critically important in the system for determining the stream's potential in terms of stability, habitat, and the associated biological potential.

Stream Channel Stability

Because streams and rivers are dynamic systems, the concept of stream-channel stability may seem hypothetical because channels naturally adjust over time. While streams are

Table 10.1. Valley types used in the geomorphic characterization and associated stream types (derived from Rosgen, 1996, 2006)

Valley types	Summary description of valley types	Stream types[a,b]
I	**Steep Canyons**: Steep, confined, V-notched canyons with rejuvenated side-slopes	**Aa+**, **A**, **G**
II	**Colluvial**: Moderately steep, gentle-sloping side-slopes often in colluvial valleys	**B**, [G]
III	**Alluvial Fans and Debris Cones**: Primarily depositional with characteristic debris-colluvial or alluvial fan landforms (a) Active alluvial fan (b) Inactive alluvial fan	A, B, [F], [G], **D**
IV	**Canyons and Gorges**: Canyons, gorges, and confined alluvial and bedrock-controlled valleys with gentle valley slopes	C, **F**
V	**Glacial Troughs**: Moderately steep, U-shaped glacial-trough valleys	**C**, **D**, F, **G**
VI	**Bedrock-Controlled**: Moderately steep, fault-, joint- or bedrock-controlled valleys	Aa+, A, **B**, C, F, [G]
VII	**Fluvial-Dissected**: Steep, fluvial-dissected, high-drainage density alluvial slopes	Aa+, **A**, F, **G**
VIII	**Alluvial**: Alluvial valley fills either narrow or wide with moderate to gentle valley slopes, well-developed floodplains, and river terraces, glacial terraces or colluvial slopes adjacent to the alluvial valley (a) **Gulch Fill Valley**: Narrow with relatively steep valley slopes (>4%) (b) **Alluvial Fill Valley**: Moderate valley widths and moderately steep valley slopes (2–4%) (c) **Terraced Alluvial Valley**: Wide valley width with gentle gradients less than 2%	**C**, [D], **E**, [F], [G], B
IX	**Glacial Outwash and Dunes**: Broad, moderate-to-gentle slopes associated with glacial outwash or eolian sand dunes	**C**, **D**, [F]
X	**Lacustrine**: Very broad and gentle valley slopes associated with glaciolacustrine and non-glaciolacustrine deposits	**C**, **DA**, **E**, [F], [G]
XI	**Deltas**: Large river deltas and tidal flats constructed of fine alluvial materials from riverine and estuarine depositional processes; most often distributary channels, wave- or tide-dominated	C, **D**, **DA**, E

[a] Bolded stream types indicate the most prevalent stream type for that valley type.
[b] Bracketed stream types are most often observed under disequilibrium conditions.

constantly adjusting, they do assume features and form that reflect their natural setting and habitat. In this regard, stream classification provides a systematic approach that furnishes reproducible and consistent results with respect to some established reference so that streams can be evaluated with one another over time. Departures from reference conditions due to people's interventions can provide indicators that the stream channel has departed from the reference condition. The challenge, therefore, is to develop some type of framework in which stream-channel stability can be evaluated.

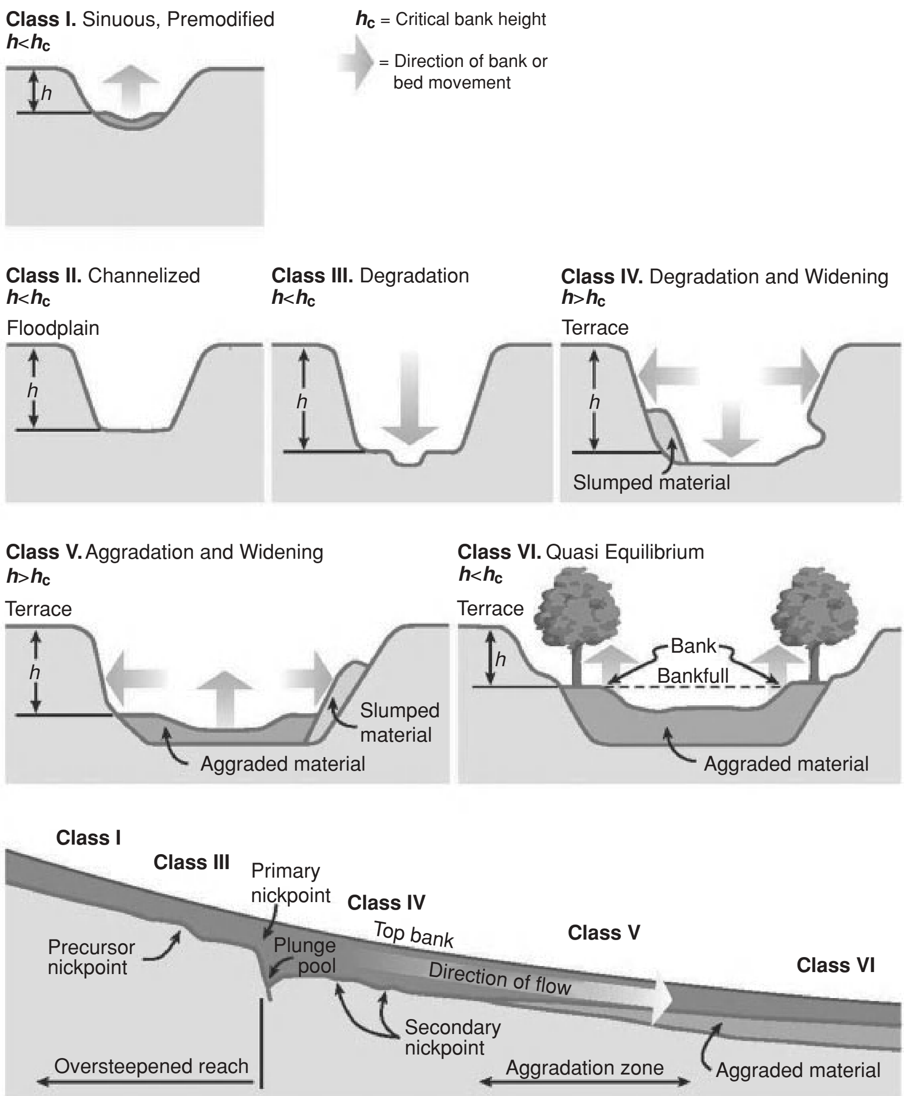

FIGURE 10.4. Channel Evolution Model moves from stable to degraded to aggregated at a new base elevation. Height *h* influences potential stream energy (modified from Simon and Hupp (1986) as presented by Simon and Rinaldi (2000) and the Federal Interagency Stream Restoration Working Group (1998) website)

Schumm's Evolution Model

A channel-evolution model was developed by Schumm et al. (1984) that provides a basis for studying the stability of alluvial channels in western Tennessee (Fig. 10.4). In this model, alluvial channels that become destabilized by people-induced or natural disturbances are shown to progress through a sequence of channel changes with time (Simon and Hupp, 1986; Simon and Rinaldi, 2000). The effect of riparian vegetation or its removal will in-turn affect channel resistance to flow and shear forces applied to a streambank. Beginning with a stable channel (Stage I in Fig. 10.4), removal of riparian vegetation, increased stream discharges, or a change in sediment supply or size will cause a channel to enlarge typically by downcutting. As the channel deepens, banks become steepened, slope failures begin to occur, and riparian trees begin to tilt toward the channel (Stages II and III). Channel banks are now destabilized and mass wasting occurs within these stages.

As soil material fills into the channel, the stream channel widens (Stage IV) and begins aggradation (Stage V). Degradation proceeds upstream as the reduced gradient in this section can no longer transport sediment from the upstream-degraded channel. As riparian vegetation becomes reestablished, the banks eventually flatten, and the hydraulic gradient is reduced through meander extension and elongation, the channel achieves a new dynamic equilibrium (Stage VI). At this point, a smaller active floodplain is formed and the old floodplain becomes a terrace. Under this new situation, flood flows are constrained by the boundaries of the new terraces that can result in higher velocities and greater erosive power during less frequent floods. These unstable channels can cause channel instability (Box 10.1) and structural failure of culverts, bridges, or road crossings.

Bank-Height Ratio

A rapid way to determine whether a channel is incising to a new form and type is through the use of the Bank-Height Ratio (BHR). The BHR is a measure of the degree of channel incision leading toward entrenchment; that is, the vertical containment of a river within a valley. Streams do not normally incise (downcut) unless the width-to-depth ratio is less than 10 (Rosgen, 2006). A BHR is calculated by

$$\mathrm{BHR} = \mathrm{LBH}/d_{\mathrm{bmx}} \qquad \textbf{(10.1)}$$

where LBH is the lowest bank height and d_{bmx} is the bankfull maximum depth.

If BHR $= 1$ and equal to the bankfull stage, the channel is considered stable and flood flows spill out of the channel onto the floodplain and possibly drop sediment because of excess sediment supply or lack of sediment transport capacity. If BHR > 1, flows greater than the bankfull discharge are required to exceed the top of the bank. A pattern of increasing or decreasing BHRs can emerge by measuring the BHR at several locations along a channel profile. For example, if the BHR increases in the downstream direction, then channel incision is advancing headward to indicate the possible development of a *knickzone*. A knickzone is similar to a knickpoint but it is transitional versus abrupt; the incision occurs over a longer distance usually forming a wedge. A wedge can be observed by looking at the bank profile from the opposite bank. If the wedge is decreasing in the downstream direction, the source(s) of channel incision is from upstream – typically a change in upland hydrology. The BHR is a useful field measurement to quickly estimate stream-channel stability and the direction of disturbance (Rosgen, 2006).

Pfankuch's Procedure for Evaluating Stream Channel Stability

Pfankuch (1975) presented a procedure for inventorying and then evaluating the stability of mountainous streams. With this procedure, several indicators are used in different positions

Box 10.1

Channel Instability: Loess Area of the Midwestern USA (Simon and Rinaldi, 2000)
Unstable stream channels are common in the loess area of the Midwest as a result of human activities on watersheds and in stream channels. This area extends south to north from Mississippi to Minnesota and Wisconsin and from east to west from Ohio to Kansas. Since 1910, channels have been enlarged and straightened to reduce flooding along their floodplains. These activities, accompanied by intensive agricultural development, urbanization, road construction, and other land-use changes on watersheds, have resulted in system-wide channel instability. Channel degradation has caused more than $1.1 billion damages to bridges, pipelines, and adjacent lands since the early 1900s. Channels have become "canyon-like" in some areas, while in others they have over-widened as a result of accelerated soil mass movement. In western Iowa, nearly 95% of the stream channels were considered unstable (Stages II–V as illustrated in Fig. 10.5).

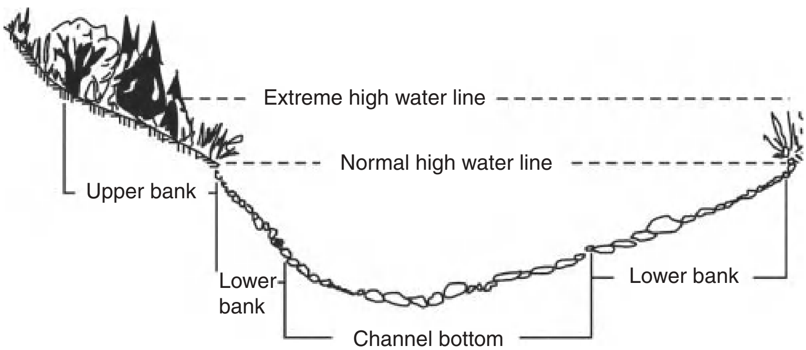

FIGURE 10.5. Stream channel and bank locations used in the Pfankuch (1975) method of streambank stability rating

Land-use activities that have contributed to channel instability include

1. Clearing of the land for cultivation, which was accelerated following the Civil War, and which increased runoff, peakflow discharges, erosion of uplands, and gullying of the landscape.
2. The removal of woody vegetation along streambanks and riparian areas, including floodplains, which reduced the hydraulic roughness, increased flow velocities, and stream power to erode streambeds and banks and transport sediment.

3. Channelization projects that dredged and straightened channels which caused streams to become entrenched. Channel degradation is widespread in the loess area, but differences in channel instability were pronounced between streams in sand beds and those in silt beds. Silt-bed streams downcut for long periods of time, as much as 70 years in western Iowa. Without coarse-grained sediment for downstream aggradation, energy is dissipated through channel widening. In the silt-bed streams, channel widening is sustained by the high streambanks and low cohesive strength of soils. Where streams cut into sands, aggradation occurs more quickly and channel widening occurs at more moderate rates. Mass wasting into stream channels is a dominant process throughout the region and extends meanders. Some riverside residents with homes and property on bluffs overlooking rivers have become victims of this process.

in a channel cross section to determine the resistive capacity of streams to the detachment of bed and bank material and provide information on the capacity of streams to adjust and recover from changes in flow, increase in sediment production, or both.

A rating score is assigned to different indicators in the channel bottom, the lower bank (the top of which is bankfull), and the upper bank (Fig. 10.5) with lower scores indicating greater streambank stability. The rating form and numerical scores for each indicator are presented in Table 10.2.

Each channel segment is rated excellent, good, fair, or poor based on the total numerical score obtained through interpretations of Table 10.2. Although subjective, these categories provide watershed managers with a means to set priorities. A stream reach with a poor rating should get more immediate attention for upstream watershed management and/or channel modifications than would one rated as good or excellent.

Minnesota Agricultural Ditch Reach Assessment for Stability

Magner et al. (2012) developed a rating system of stream stability for ditches that are dug into flat and poorly drained landscapes with isolated basins in the corn belt region of the Midwestern USA. Building on the CEM discussed above, the Minnesota Agricultural Ditch Reach Assessment for Stability (MADRAS) system considers factors associated with channelized water, groundwater seepage, and sediment transport. For example, many ditches in the corn belt region are cut through drained wetlands. Wetland soils hold shallow groundwater unless drained with plastic pipe, also referred to as tile drains. If a ditch bank becomes steep and the water table is not controlled by tile drains, water can seep through the ditch bank causing the bank to become unstable and collapse into the ditch bed. This rating system is easily implemented if the evaluator has basic fluvial knowledge and the ability to detect groundwater seepage. However, as the case with any observational tool,

TABLE 10.2. Stream reach stability evaluation

	Key no.	Stability indicators by classes: Fair		Poor	
Upper banks	1	Bank slope gradient 40–60%	(6)	Bank slope gradient 60% +	(8)
	2	Moderate frequency and size, with some raw spots eroded by water during high flows	(9)	Frequent or large, causing sediment nearly yearlong or imminent danger of same	(12)
	3	Present volume and size are both increasing	(6)	Moderate to heavy amounts, predominantly larger sizes	(8)
	4	50–70% density; lower vigor and still fewer species form a somewhat shallow and discontinuous root mass	(9)	<50% density plus fewer species and less vigor indicate poor, discontinuous, and shallow root mass	(12)
Lower banks	5	Barely contains present peaks; occasional overbank floods; W/D ratio 15–25	(3)	Inadequate; overbank flows common; W/D ratio >25	(4)
	6	20–40%, with most in the 3–6″ diameter class	(6)	Frequent obstructions and deflectors cause bank erosion yearlong; sediment traps full, channel migration occurring	(8)
	7	Moderately frequent, moderately unstable obstructions and deflectors move with high water causing bank cutting and filling of pools	(6)	Frequent obstructions and deflectors cause bank erosion yearlong; sediment traps full, channel migration occurring	(8)
	8	Significant; cuts 12–24″ high; root mat overhangs and sloughing evident	(12)	Almost continuous cuts, some over 24″ high; failure of overhangs frequent	(16)
	9	Moderate deposition of new gravel and coarse sand on old and some new bars	(12)	Extensive deposits of predominantly fine particles; accelerated bar development	(16)
Bottom	10	Corners and edges well rounded in two dimensions	(3)	Well rounded in all dimensions, surfaces smooth	(4)
	11	Mixture, 50–50% dull and bright, ± 1%, i.e., 35–6%	(3)	Predominantly bright, 65% +, exposed or scoured surfaces	(4)
	12	Mostly a loose assortment with no apparent overlap	(6)	No packing evident; loose assortment, easily moved	(8)
	13	Moderate change in sizes; stable materials 20–50%	(12)	Marked distribution change; stable materials 0–20%	(16)
	14	30–50% affected; deposits and scour at obstructions, constrictions, and bends; some filling of pools	(18)	More than 50% of the bottom in a state of flux or change nearly yearlong	(24)
	15	Present but spotty, mostly in backwater areas; seasonal blooms make rocks slick	(3)	Perennial types scarce or absent; yellow-green, short-term bloom may be present	(4)
			Total		Total

Table 10.2. (*Continued*)

	Key no.	Stability indicators by classes			
		Excellent		**Good**	
Upper banks	1	Bank slope gradient <30%	(2)	Bank slope gradient 30–40%	(4)
	2	No evidence of past or any potential for future mass wasting into channel	(3)	Infrequent and/or very small; mostly healed over; low future potential	(6)
	3	Essentially absent from immediate channel area	(2)	Present but mostly small twigs and limbs	(4)
	4	90% + plant density; vigor and variety suggests a deep, dense, soil binding, root mass	(3)	70–90% density; fewer plant species or lower vigor suggests a less dense or deep root mass	(6)
Lower banks	5	Ample for present plus some increases; peak flows contained; W/D ratio <7	(1)	Adequate; overbank flows rate; W/D ratio 8–15	(2)
	6	65% + with large, angular boulders +12″ numerous	(2)	40–65%, mostly small boulders to cobbles 6–12″	(4)
	7	Rocks and old logs firmly embedded; flow pattern without cutting or deposition; pools and riffles stable	(2)	Some present, causing erosive crosscurrents and minor pool filling; obstructions and deflectors newer and less firm	(4)
	8	Little or none evident; infrequent raw banks less than 6″ high generally	(4)	Some, intermittently at outcurves and constrictions; raw banks may be up to 12″	(6)
	9	Little or no enlargement of channel or point bars	(4)	Some new increase in bar formation, mostly from coarse gravels	(8)
Bottom	10	Sharp edges and corners, plane surfaces roughened	(1)	Rounded corners and edges, surfaces smooth and flat	(2)
	11	Surfaces dull, darkened, or stained; generally not “bright”	(1)	Mostly dull, but may have up to 35% bright surfaces	(2)
	12	Assorted sizes tightly packed and/or overlapping	(2)	Moderately packed with some overlapping	(4)
	13	No change in sizes evident; stable materials 80–100%	(4)	Distribution shift slight; stable materials 50–80%	(8)
	14	Less than 5% of the bottom affected by scouring and deposition	(6)	5–30% affected; scour at constrictions and where grades steepen; some deposition in pools	(12)
	15	Abundant; growth largely moss-like, dark green, perennial; in swift water too	(1)	Common; algal forms in low velocity and pool areas; moss here too and swifter waters	(2)
			Total		Total

Source: From Pfankuch (1975).
Stability ratings: Add column scores – total score <38 = excellent; 39–76 = good; 77–114 = fair; 115+ = poor.

MADRAS is subjective with the exception of measuring the bank angles. MADRAS differs from the CEM because the CEM is based on the assumptions that

- there will be an excessive mobile sediment supply (sand and gravel) from the watershed; and
- successive stages of evolution are driven by threshold land-use changes and not interrupted by other successive land-use disturbances that alter the contributing drainage area; for example, intensive drainage development.

Much of the landscape of the Midwestern USA is disturbed almost every year by farming operations or where surface- or subsurface-tile drains are installed before planting crops to minimize the crop failure. Field management can alter hydrologic pathways, setting into motion new channel adjustments. Therefore, ditches can be at varying stages of channel evolution based on land-use change. Based on field sheets developed by Simon and Downs (1995) for their Rapid Geomorphic Assessment (RGA) method, MADRAS adapts some of the RGA metrics to a field assessment form (Table 10.3) for ditches with flat gradients, cohesive soils, and potential groundwater discharge.

Both the Pfankuch (1975) and MADRAS stability ratings are useful in evaluating possible needs for improvements at culvert and bridge sites or where livestock are concentrated near channels. Pfankuch's method has been incorporated into the Rosgen's stream classification and assessment approach described in the following section.

Watershed Assessment of River Stability and Sediment Supply

Another evaluation tool that considers more than channel processes and explores the relationships between landscape settings in a watershed and the channel response to management practices is Watershed Assessment of River Stability and Sediment Supply (WARSSS). WARSSS was developed by Rosgen (2006) for the US Environmental Protection Agency to provide estimates of Total Maximum Daily Loads (TMDLs) of sediment across the USA. The method has three main components:

1. Reconnaissance Level Assessment (RLA)
2. Rapid Resource Inventory for Sediment and Stability Consequence (RRISSC)
3. Predictive Level Assessment (PLA)

The RLA examines existing information that considers upland and hillslope processes and ranks the relative importance of subwatersheds for sediment supply and delivery. The RRISSC level focuses on the highest priority subwatersheds that are considered at risk of future degradation or have degraded and might require restoration. This procedure requires validation of existing data from government sources and the field collection of additional watershed information such as land use and a landscape inventory followed by a rating of hillslope, hydrologic, and channel processes. Figure 10.6 illustrates the flow of analysis for the RRISSC procedure that leads to the most important locations for PLA work within a watershed.

TABLE 10.3. Form for applying the Minnesota Agricultural Ditch Reach Assessment for Stability (MADRAS) system (Magner et al., 2012)

Worksheet Metrics and Scores:

		Percent of area			
		0%	**1–10%**	**10–50%**	**>50%**
Left bank condition	Slumpage*	1	2	4	8
Right bank condition	Slumpage*	1	2	4	8
Bed deposition*		1	2	6	12
	Supportive evidence				
Channel slope*	Riffle*	Run*	Pooled*	Slow moving*	Backed up*
	1	1	2	4	8
Bank angle*		45° stable	2-angles*	90° unstable	Seepage*
		1	2	6	8
Scour*		None	Present	Toe undercut*	Extreme
		1	2	6	8
	Type, age, and condition				
Crossing*	Old culvert Small size* Rusted	Old bridge Constriction* Wooden	Perched Culvert* Any age	Sediment* Culvert New	Sediment* Bridge New
	1	1	4	8	8

Worksheet Metric Definitions:

Slumpage* = mass failure of bank material that has fallen into the channel bed
Bed deposition* = too much sediment accumulating on channel bed
Channel slope* = measured fall of water over a defined distance
Riffle = steep slope over a short distance with stream power to move water and sediment
Run = moderate slope with weak to adequate water and sediment transport
Pooled = little or no slope, deep water >2 feet, leading to saturated bank soil
Slow moving = little or no slope, shallow water (<1 foot) that appears stagnant
Backed up = suggests aggradation and loss of low-flow sediment transport
(Beaver activity or a debris jam will cause the water level to rise)
Bank angle* = direction of change (stable-to-unstable)
2-angles* = a continuous angle does not exist, compound channel
Is any portion of the bank angle ≥90°? If so rank it unstable
Seepage* = evidence of groundwater seeping out of a channel bank
Scour* = opposite of deposition, eating the channel that leads to enlargement
Is there any evidence that scouring is present? If so rank at 2
Toe Undercut* = is the toe slope of the bank exposed (bare soil)?
Extreme = abundant evidence of scour at several locations
Crossing* = the type, relative age and observable features
Old culvert/bridge = may suggest flow stability, thus channel stability
Perched culvert = suggests channel downcutting loss of vertical stability
Sediment = overwide culvert or bridge allows for slow velocity and deposition
(a scour hole below crossing suggests a change hydrology)

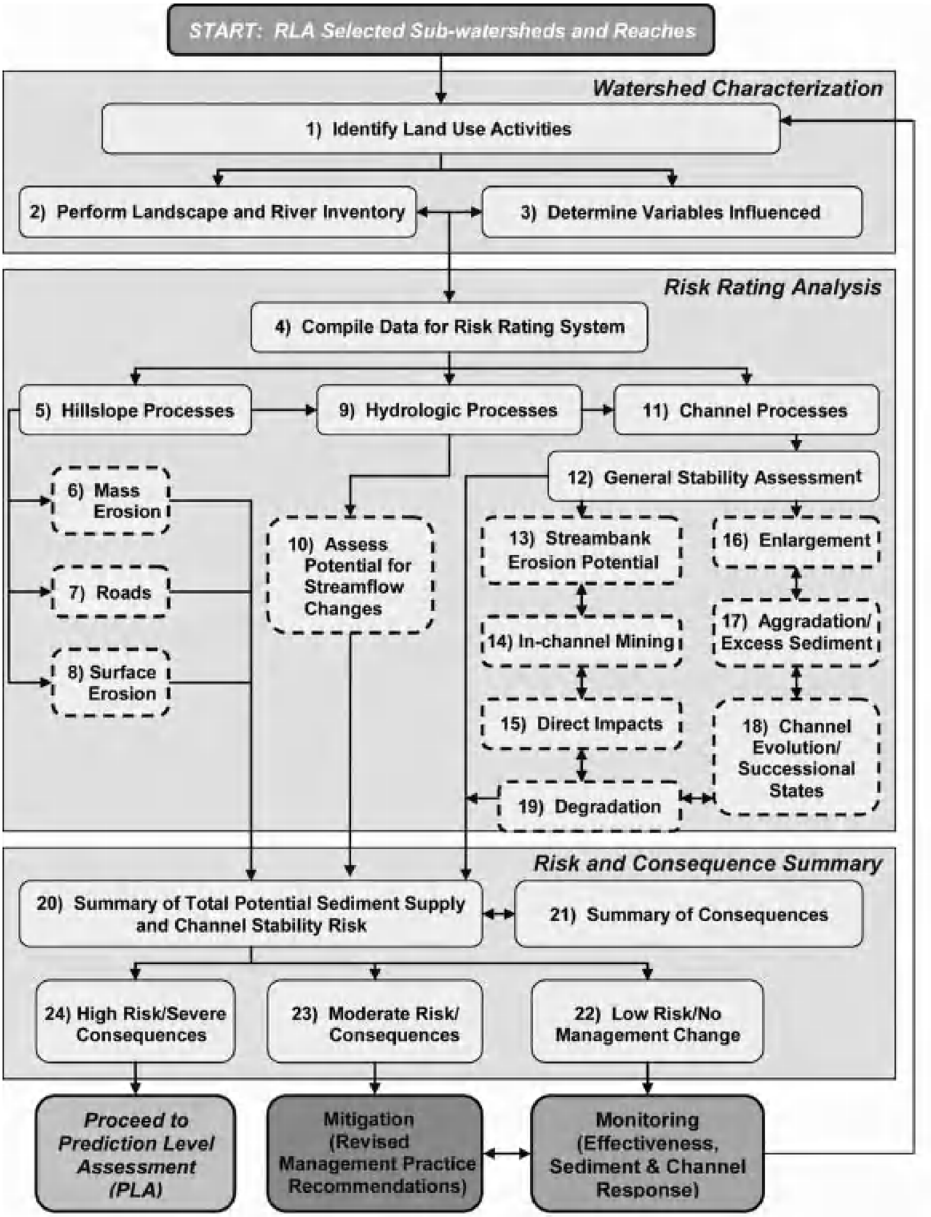

FIGURE 10.6. Flowchart 4-1 FROM WARSSS illustrating sequential steps taken in the RLA selection process for subwatersheds and reaches that required further investigation involving field data collection (from Rosgen (2006). © Wildland Hydrology, with permission)

The PLA is the most detailed component of the WARSSS methodology. This component satisfies the following objectives:

- to locate and quantify sediment sources;
- to link sediment sources to land uses;
- to identify disproportionate sediment supplies;
- to evaluate sediment impacts on stream channels;
- to integrate hydrology, river morphology, and stability with land-use impacts by location;
- to determine departure and degree of impairment due to sediment sources, watershed hydrology, and riparian impacts; and
- to provide necessary details to design site- and process-specific mitigation.

An understanding of stream classification is required to apply WARSSS or the above-channel evaluation procedures. The next section presents information necessary to conduct the WARSSS method.

Stream Classification

Stream systems are complex. To understand these systems and place them into a management framework requires knowledge of the processes that influence the pattern and character of the systems (Rosgen, 1994). However, there is often more information available on stream systems than can be efficiently applied in a watershed management context. Part of the problem confronting watershed managers in this regard is the large number of component pieces making up the data sets describing the stream form and fluvial processes involved. Definitions of the terminology used in stream classification are presented in Table 10.4.

Table 10.4. Terminology used in describing stream morphology and classification

Term	Definition
Anastomosis	A stream channel with connections between the parts of its branching system
Bankfull stage	A stage where water completely fills the channel system of a stream without spreading onto the floodplain; flow at this stage is often assumed to control the form of alluvial stream channels and is associated with a recurrence interval averaging 1.5 years
Braided	Multiple stream channels woven together
Meander	A stream channel with a winding or indirect course
Pool	A section of a stream channel with deeper, slow-moving water with fine bed material
Riffle	A section of a stream channel with shallow depth and fast-moving water; bed materials are coarse
Sinuosity	A measure of the number of bends, curves, and meanders in a stream channel compared to a straight channel (channel length/straight-line distance from upstream to downstream point)
Sinuous	A stream channel of many curves, bends, or turns; a stream channel which is intermediate between straight and meandering
Thalweg	The deepest part of the channel

Stream classification is one way in which these "pieces" of data can be brought together with the disciplines represented into a common and usable format. Several methods of classifying streams and rivers have been developed based on hydrologic and geomorphic characteristics (Gordon et al., 1992). While stream-classification scheme risks oversimplification, it is helpful in organizing the relevant data in a logical and reproducible way. Such a morphological approach has been developed for mountain streams by Montgomery and Buffington (1997). A widely applied method that has been adopted by the Stream Systems Technology Center (2001) of the US Forest Service is that of Rosgen (1994, 1996). Streams and rivers are classified in this system to

- predict a stream's behavior from its appearance;
- develop specific hydraulic and sediment relationships for a given morphological-channel type and forms;
- provide a mechanism to extrapolate site-specific data collected on a particular stream reach to those of similar character; and
- provide a consistent and reproducible frame of reference for communication among watershed managers working with stream systems in a variety of disciplines.

Stream channel width, depth, and slope, roughness of the channel materials, stream discharge and velocity, and sediment load and sediment size determine the morphology of the channel (Leopold et al., 1964). A change in one of these variables causes a series of channel adjustments that inevitably lead to changes in the others, resulting in channel pattern alteration. Because stream morphology is the product of this integrative process, the variables that are measurable should be used as stream classification criteria. This is the basis of the Rosgen's Stream Classification (RSC) described below.

Rosgen's Stream Classification

The level of classification presented by Rosgen (1994) is commensurate with many planning-level objectives in watershed management. A hierarchy of stream classification and inventories is desirable, therefore, because these objectives vary. Such a hierarchy allows an organization of stream inventory data into levels of resolution ranging from broad morphological characteristics (Level I) to reach specific descriptions (Levels II, III, and IV) as described in Table 10.5. Level I classifies the stream from type Aa+ to G. The patterns of a stream are classified as relatively straight and steep (Aa+ and A stream types), low sinuosity and moderate gradient (B stream types), to low gradient and meandering (C and E stream types). Complex multiple-channel stream patterns are the braided (D) and anastomosis (DA) stream types. Types F and G represent entrenched and often-unstable streams that characteristically exhibit high rates of bank erosion; however, the Grand Canyon in Arizona is an example of a stable F stream type. The Aa+ and A streams are steep, entrenched, straight reaches with rapidly flowing water typical of high-elevation mountainous streams. As the hydraulic gradient diminishes, streams meander and their sinuosity increases.

Level II further subdivides streams according to entrenchment ratios, slope ranges, and dominant-channel material-particle sizes. The *entrenchment ratio* is the ratio of the width of the flood-prone area to the bankfull surface width of the channel. The flood-prone area is determined from the flood-prone depth that is twice the maximum bankfull depth. The stream types are assigned numbers related to the median diameter of particles: 1 = bedrock, 2 = boulder, 3 = cobble, 4 = gravel, 5 = sand, and 6 = silt/clay. These separations

TABLE 10.5. Levels of the Rosgen Stream Classification system

Level	Description of inventory	Objectives	Information required
I	Broad morphological classification	For generalized description of major types of streams	Landform, lithology, soils, climate, basin relief, deposition history, valley morphology, river profile morphology and pattern
II	Morphological description of stream types	Provide delineation and detailed interpretation of homogeneous stream reaches based on reference reach measurements	Channel patterns, entrenchment ratio, width/depth ratio, sinuosity, channel material, slope
III	Stream state or condition	Determine existing condition or stability of streams; estimate departure from their potential	Channel stability index, riparian vegetation, deposition patterns, debris occurrence, bank erosion, confinement features, fish habitat indices, river size category
IV	Verification	Verify the stream condition, stability, and potential as predicted from Levels I, II, and III	Streamflow, suspended sediment and bedload measurements, bank erosion rates, aggradation–degradation of channel beds, biological data, riparian vegetation measurements, hydraulic geometry

Source: From Rosgen (1994, 1996).

initially produce 41 major stream types as shown in Figure 10.7. A range of values for each criterion are presented in the key to classification for the 41 major stream types (Fig. 10.8). The values selected for each criterion were obtained from data sets obtained from a large assortment of streams throughout the USA, Canada, and New Zealand.

The RSC is applicable for a variety of conditions. It can be applied to either ephemeral or perennial channels with little modification. An important feature in this classification is the ability to identify the bankfull stage in the field. For many streams, bankfull stage is the level where water completely fills the stream channel, in such cases it can be identified in most perennial channels with field indicators. These bankfull indicators are often more elusive in ephemeral channels. Although originally developed for application in the streams of the Rocky Mountains, the RSC has also been successfully applied in low-relief terrain and low-gradient stream channels in Wisconsin (Savery et al., 2001), moderate-gradient streams in clay-bed channels in Minnesota (Riedel, 2000), and elsewhere. Regional relationships of bankfull stage and watershed area have been developed in some regions of Minnesota (Magner and Brooks, 2007).

The morphological variables considered in the classification often change in short distances along a stream channel largely due to changes in geology and the tributaries. As a consequence, the morphological descriptions are based on observations and measurements made on representative reaches of the stream from only a few meters to several kilometers. The stream types classified by the system apply only to these selected reaches of the channel.

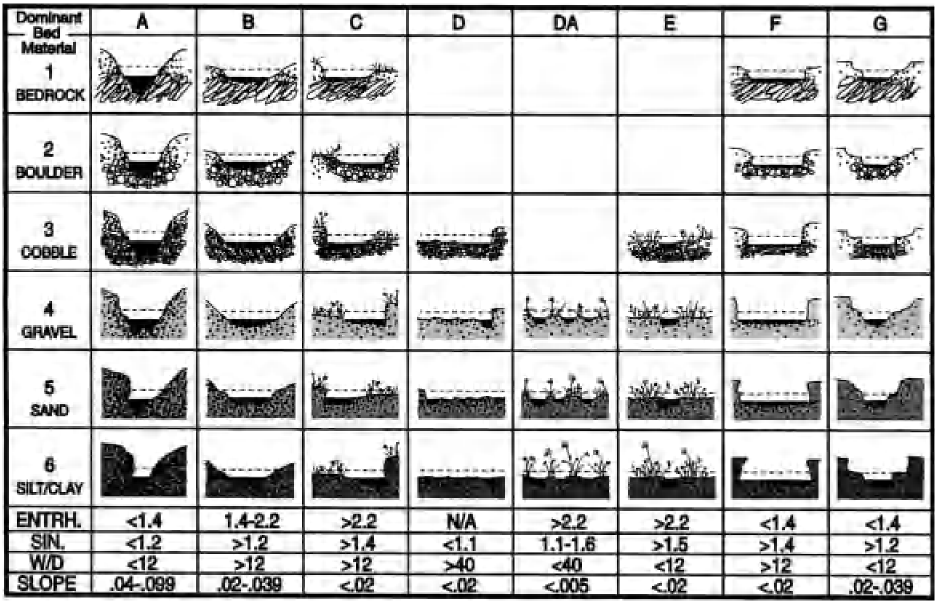

FIGURE 10.7. Major stream types with their cross-sectional configurations and physical characteristics (from Rosgen (1994). © Catena, with permission)

The observations and measurements from selected reaches are not averaged over the entire watershed to classify stream systems. The stream classifications obtained correspond with the valley types discussed above.

Continuum Concept

The RSC recognizes a continuum of stream morphology within and between stream types. This continuum is applied where values outside the "normal range" are encountered but do not warrant classification of the stream as a unique type. The general dimensions and patterns of a stream do not generally change with a minor change in one of the delineative criteria of the system.

The ranges in slope, width/depth ratio, entrenchment ratio, and sinuosity shown in Figure 10.7 span the most commonly observed field-derived values. Exceptions in which the value of one of these variables is outside the range for a given stream type occur infrequently.

Streams do not usually change instantaneously but more frequently undergo a series of channel adjustments through time to accommodate any change in the "driving" variables. Their dimensions, profiles, and patterns reflect these adjustment processes, which are largely responsible for the observed stream form. The rate and direction of channel adjustment are functions of the nature and magnitude of the change, valley type, and the stream type involved. Some streams change rapidly due to land-use change while others are slow in their response to change.

Processes of degradation and aggradation (see Chapter 9) can cause a change in stream type. For example, aggradation can trigger changes in width/depth ratio, slope, and

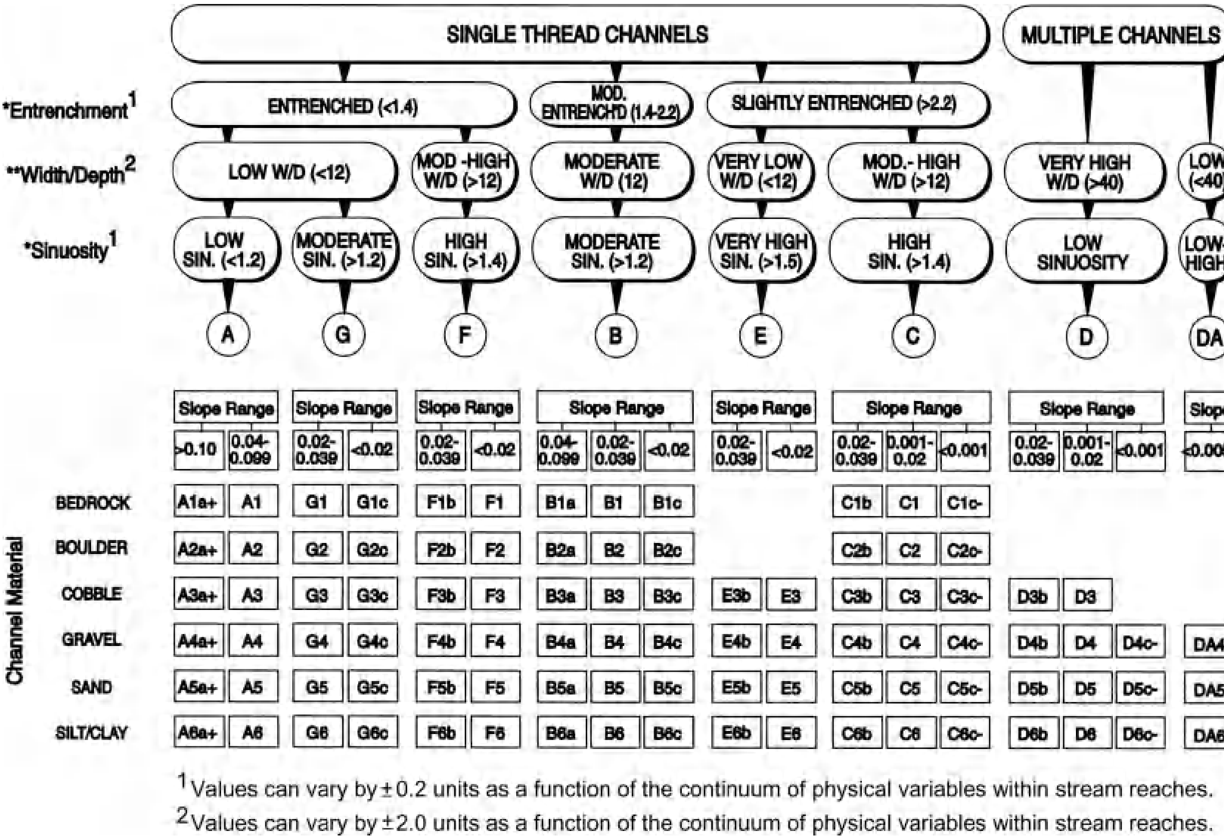

FIGURE 10.8. Classification system for natural rivers (from Rosgen (1994). © Catena, with permission)

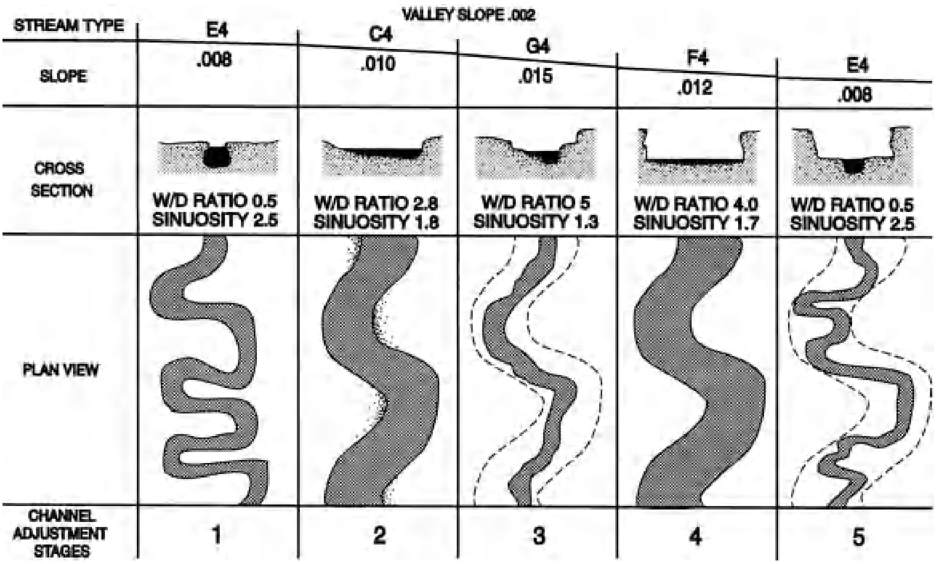

FIGURE 10.9. Example of evolutionary adjustments in stream channels (from Rosgen (1994). © Catena, with permission)

sinuosity; type E streams can change to type C (Fig. 10.9). Conversely, degradation would involve moving from C to G or E to G type streams.

Stream Channel Succession

The information provided by Rosgen (2006) coupled with earlier criticism about the RSC being a form-based approach that was not process oriented lead to the creation of the Rosgen's Stream Channel Succession (RSCS) models. Figure 10.5 illustrated the concept of channel evolution described by Simon and Hupp (1986). Figure 10.10 depicts the same steps in the CEM and also illustrates four separate Rosgen stream types that approximately capture the processes described by the CEM. An additional 11 RSCS models have been developed to describe processes that have been observed (Schumm, 2005), but difficult to communicate without a well-developed stream-classification system. Although the CEM provides a broad approach, the RSCS provides greater detail that can be useful for resource managers.

Interpretations of Stream Classification

The ability to predict the behavior of a stream from its appearance and to extrapolate information from similar stream types is helpful to watershed managers in applying interpretive information. These interpretations can help a watershed manager to evaluate stream types in relation to their sensitivity to disturbance, recovery potential, sediment supply, the controlling influence of vegetation, and streambank erosion potential (Box 10.2). Applications of these interpretations can be useful in impact assessment, risk analysis, and management direction by stream type.

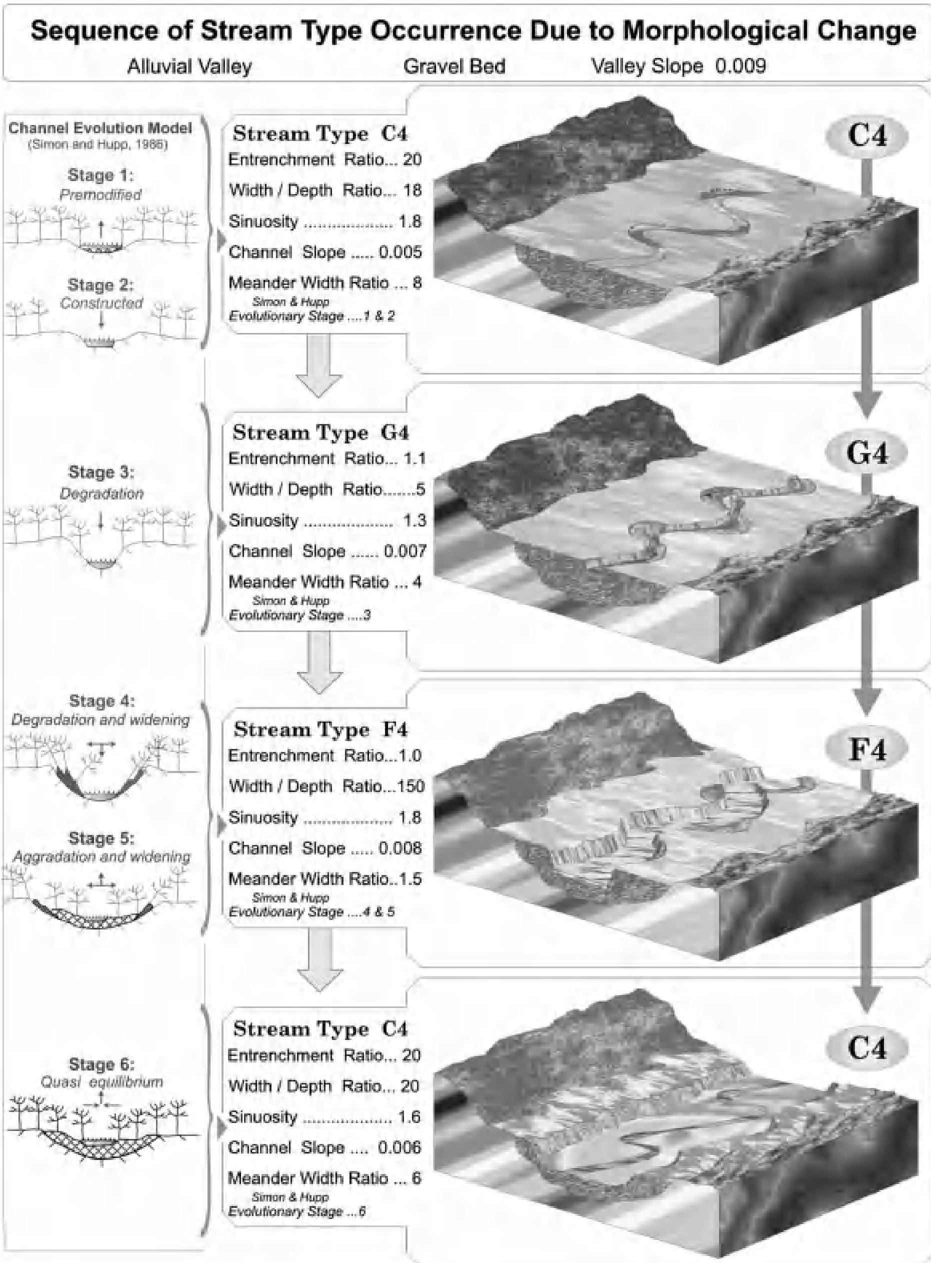

FIGURE 10.10. Comparison of the Channel Evolution Model (from Simon and Hupp, 1986) and the corresponding four Rosgen stream types (from Rosgen, 2006. © Wildland Hydrology, with permission) (For a color version of this photo, see the color plate section)

Box 10.2

Management Interpretations of Stream Types Classified by Rosgen (1994)

Stream type	Sensitivity to disturbance[a]	Recovery potential[b]	Sediment supply[c]	Streambank erosion potential	Vegetation controlling influence[d]
A1	Very low	Excellent	Very low	Very low	Negligible
A3	Very high	Very poor	Very high	High	Negligible
A4	Extreme	Very poor	Very high	Very high	Negligible
A6	High	Poor	High	High	Negligible
B1	Very low	Excellent	Very low	Very low	Negligible
B3	Low	Excellent	Low	Low	Moderate
B4	Moderate	Excellent	Moderate	Low	Moderate
B6	Moderate	Excellent	Moderate	Low	Moderate
C1	Low	Very good	Very low	Low	Moderate
C3	Moderate	Good	Moderate	Moderate	Very high
C4	Very high	Good	High	Very high	Very high
C5	Very high	Fair	Very high	Very high	Very high
C6	Very high	Good	High	High	Very high
D3	Very high	Poor	Very high	Very high	Moderate
D5	Very high	Poor	Very high	Very high	Moderate
D6	High	Poor	High	High	Moderate
DA4	Moderate	Good	Very low	Low	Very high
DA6	Moderate	Good	Very low	Very low	Very high
E3	High	Good	Low	Moderate	Very high
E4	Very high	Good	Moderate	High	Very high
E6	Very high	Good	Low	Moderate	Very high
F1	Low	Fair	Low	Moderate	Low
F3	Moderate	Poor	Very high	Very high	Moderate
F4	Extreme	Poor	Very high	Very high	Moderate
F6	Very high	Fair	High	Very high	Moderate
G1	Low	Good	Low	Low	Low
G2	Moderate	Fair	Moderate	Moderate	Low
G3	Very high	Poor	Very high	Very high	High
G4	Extreme	Very poor	Very high	Very high	High
G6	Very high	Poor	High	High	High

[a] Includes increases in streamflow magnitude and timing and/or sediment increases.
[b] Assumes natural recovery once cause of instability is corrected.
[c] Includes suspended and bed load from channel-derived sources and/or from stream-adjacent slopes.
[d] Vegetation that influences width/depth ratio stability.

Levels III and IV provide more detailed descriptions and characteristics of stream channels that provide indications of stream potential, stability, and the status of stream channels in comparison with their potential. Such detail is required to understand and link past land use with channel changes and to help guide stream restoration efforts.

Interpretive information by stream type can also be used in establishing guidelines for watershed management practices, silvicultural standards, and riparian and floodplain management and in analyzing possible cumulative effects.

Summary and Learning Points

Energy relationships of flowing water and hydraulic principles help to explain geomorphic processes that determine valley type, stream-channel form, and function. An understanding of these relationships and linkages is crucial to achieving effective watershed management. This chapter considers these linkages and the application of geomorphic stream-classification systems to understand how hydrologic changes and cumulative land-use effects can alter stream channels. After reading this chapter, you should be able to

1. Explain hydrologic functions and processes that influence fluvial geomorphology and the resulting landscape features such as valleys, terraces, floodplains, and the bankfull flow condition.
2. Discuss the relevance of bankfull or channel forming flows in terms of stream channel morphology.
3. Describe factors that determine the form of a stream channel and indicate how changes in streamflow can affect stream channel form, function, stability and evolution or succession over time.
4. Recognize the importance of valley and stream classification in helping to predict a stream's behavior from its appearance.
5. Explain the role of stream classification in establishing guidelines for watershed management, riparian management, and channel management in analyzing cumulative effects and restoration needs.

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CHAPTER 11 Water-Quality Characteristics

INTRODUCTION

Before we can discuss water quality, we must identify how the water is to be used. Water quality involves a long list of individual components and physical, chemical, and biological constituents. A *water-quality standard* refers to the physical, chemical, or biological characteristics of water in relation to a specified use or designated use defined by government numerically or by narrative. For example, water-quality standards for irrigation are not necessarily acceptable for drinking water. Changes in water quality resulting from land-use activities on a watershed can make water unusable for drinking but it can still be acceptable for irrigation or other uses. In some instances, we are required by laws or regulations to prevent water-quality characteristics from being degraded from natural or background conditions defined as *nondegradation*, and, as a consequence, these characteristics are a key provision in the water-quality standards. The objective of these laws and regulations is to maintain the quality of water for a possible unforeseen use in the future use (see Chapter 14).

The term *pollution* refers to water that has been degraded or defiled in some way by people's actions. Violation of water-quality standards can result from nonhuman actions; however, the standard should be adjusted to account for natural background characteristics. It should be recognized that failure to meet water-quality standards occurs from natural phenomena such as fine-lake-clay sediment and events such as large rainstorms, wildfires, or volcanic eruptions. Pollution that is the result of land use that is poorly planned or implemented is the focus of this discussion. In the United States, the primary tool being applied to manage pollution is known as the total maximum daily load (*TMDL*) that focuses on the quality of water in streams and lakes rather than an end-of-pipe effluent standard.

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

End-of-pipe refers to point sources of pollution that discharge treated water into a surface water body but do not account for nonpoint-source pollution.

This chapter deals with water-quality characteristics of naturally occurring surface water in nonurban, rural, or wildland settings and land-use impacts on these characteristics. However, surface and subsurface water can often exchange (see Chapter 7). As a consequence, the source of pollution requires a whole watershed systems thinking approach. The discussion focuses on nonpoint-source rather than point-source pollution; however, the fundamentals of point-source pollution, particularly oxygen demand is addressed in this chapter. *Point-source pollution* is associated with industries or municipalities and the discharge of pollutants through a pipe or ditch into designated surface water. *Nonpoint-source pollution* (*NPS*) (Magner, 2011) refers to pollution that is associated with land-use activities such as agricultural cultivation, grazing of livestock, and forest management practices which can occur in an imperceptible manner over wide areas. Therefore, we must consider how to monitor and assess water quality in nonurban watersheds in a way that leads to system understanding which is vital to implementing management actions.

Management responses to *NPS* pollution problems are reviewed in Chapters 12 and 13. Groundwater quality in reference to its usefulness for domestic use and irrigation is also considered in this chapter. However, we further discuss the geochemical and isotopic attributes of water as influenced by the atmosphere, rock, soil, microorganisms, and people. We begin this discussion of water-quality characteristics with a brief look at the chemistry of the precipitation that falls onto watershed lands and, therefore, influences the quality of water flowing from these lands.

Chemistry of Precipitation

Air pollution and its influence on precipitation chemistry and dry-deposition additions to terrestrial and aquatic systems are of interest to watershed managers, farmers, and society as a whole. The additions of higher concentrations of chemicals to land and water and other changes in water chemistry can affect land productivity and water quality with environmental implications globally.

Acid Precipitation and Atmospheric Chemical Constituents

Acid precipitation became a major environmental issue in the late 1970s and early 1980s in the industrial countries of the northern hemisphere. Acid refers to *pH* values less than 7 in water where *pH* is defined as the negative log (base 10) of the hydrogen (H^+)-ion activity in moles per liter (*pH* is discussed in more detail later in the chapter). This issue prompted a global program of monitoring the chemistry of precipitation. In the United States, the National Atmospheric Deposition Program (*NADP*) was created in 1977 to determine atmospheric-chemical deposition and its effects. Concentrations of sulfate (SO_4), nitrate (NO_3), ammonium (NH_4), calcium (*Ca*), and H^+ ions are measured at the *NADP* sites (Kapp et al., 1988). Such programs have provided insight into the chemical characteristics of both rainfall and snowfall.

Naturally occurring precipitation is slightly acidic (*pH* values of 5.6–5.7) because of the reaction of water with normal levels of atmospheric carbon dioxide (CO_2) (Environmental

Protection Agency, 1980). The industrial northeastern United States experienced increasing acidity in the late 1970s and early 1980s with large areas experiencing rainfall *pH* values of 4.5 and below. In northern Europe, *pH* values as low as 2.4 were reported. These values are close to the acidity of lemon juice. The increasing acidity of precipitation was caused by the atmospheric inputs of sulfur (S^-) and nitrogen (N) oxides from the burning of fossil fuels such as coal, gas, and oil. Of major concern are the impacts of acid deposition including both liquid and solid atmospheric particulates on aquatic and terrestrial ecosystems.

The effects of acid deposition are a global concern because of the long-range transport of acid by the atmosphere. The impacts of acid deposition on streams and lakes are largely a function of the buffering capacity of soil surrounding them and the size of the watershed. Acids become neutralized and waters may not become acidified if soils are alkaline or contain sufficient calcium (Ca). However, lakes and streams that occur on infertile and shallow soils on top of acidic bedrock and have a low ratio of watershed area to water-surface area are susceptible to acidification. Fish and organisms providing their food are generally affected once the *pH* of streams and lakes drops below 6.0.

Also, of importance is the impact of acid deposition on vegetation and ecological processes. Acid deposition can inhibit bacterial decomposition and N fixation in the soil and can cause Ca, magnesium (Mg), and potassium (K) to be leached from the soil. Air pollutants including acid deposition caused a \$1.2 billion loss of trees in Germany in the summer of 1983 alone (Postel, 1994). Under severe cases around smelters, forests can become nonproductive and a loss of plant cover can lead to the same serious hydrologic problems encountered with other barren landscapes.

Snowpacks in the northern hemisphere can also be acidic. Snowflakes can collect a variety of impurities by scavenging as they descend through the atmosphere. Dry-atmospheric fallout between snowstorms and local contamination can also add to the impurities found in snowpacks. Snowpack acidity was generally not a problem in the western United States and Canada with the possible exceptions of "hot spots" that were located downwind of major industrial developments (Box 11.1).

Box 11.1

Sources of Contaminants in Snowpacks – An Example in Montana, United States (Pagenkopf, 1983)

The low *pH* values of snowpacks observed in the southwest corner of Montana were initially puzzling because there is little industry or population in the vicinity. Research indicated that the source of the acid oxides originated from five distinct areas outside of Montana: the Seattle-Tacoma area of Washington, the Portland area of Oregon, the San Francisco Bay area, the greater Los Angeles basin, and the Wasatch front in Utah. Apparently, storms that generally track from west or southwest to east cross the mountains of western Montana, deposit the accumulated sulfuric and nitric acids with the falling snow.

Controlling emissions is the key to reducing acid precipitation. Acid precipitation in North America and Europe has decreased in the last two decades of the twentieth century by reducing emissions of sulfate (SO_2) and nitrate (NO_x). According to the Pacific Research Institute (http://en.wikipedia.org/wiki/Pacific_Research_Institute) from the 1990s to 2007, SO_2 emissions dropped 40% in the United States and acid-rain levels dropped 65% since 1976. The European Union has reported more than 70% reductions in SO_2 over the same period. Given these reductions, other sources of pollution such as mercury (Hg) have been identified as a more pressing issue.

Atmospheric Deposition of Mercury

One of the main sources of Hg to watersheds, streams, and lakes is by atmospheric deposition (Kolka, 1996). As a trace element in geologic formations, Hg is naturally released by the slow process of weathering. This process does not yield high concentrations of Hg to terrestrial and aquatic systems. However, there are several sources of Hg resulting from natural events and people's activities in which Hg becomes volatile. Forest fires and volcanic eruptions expel Hg to the atmosphere. People's activities including metal production, waste handling and treatment, and burning of fossil fuels, wood, and peat contribute to atmospheric Hg.

Once in the atmosphere, Hg becomes widespread and has been found in remote wilderness areas. Hg has a strong affinity for organic matter in soils and surface water. As a result, Hg transport and cycling in a watershed is closely tied to the processes and movement of organic carbon. Boreal forests and peatlands in the northern hemisphere are a source for organic carbon that leaves watersheds in this region. Atmospheric Hg apparently accumulates on the forest canopy and reaches the soil which acts as a sink. For example, studies in northern Minnesota indicate that from 50% to 80% of the Hg export was associated with particulate organic carbon in runoff from uplands and peatlands (Kolka, 1996; Kolka et al., 2001). This represents the main source of Hg that is reaching streams and rivers and affecting the aquatic food chain in this region. The risks of accelerated Hg levels to wildlife, fish, and humans are well known (Trasande et al., 2005).

Physical Characteristics of Surface Water

Among the more important physical characteristics of surface water are suspended-sediment concentrations, bed-particle size, water temperature, and the level of dissolved oxygen (DO). Suspended sediments, which consist largely of silts and colloids of various materials, affect water quality for domestic and industrial uses and can adversely affect aquatic organisms and their environments such as the loss of fish-spawning habitat via embeddedness. Thermal pollution also has many direct and indirect impacts on aquatic organisms and water quality. While DO is a physical measure of water quality, it is also an important determinant of chemical and biological processes in water. Electrical conductivity (EC) of water or the temperature-corrected (to 25°C) measure of conductivity known as specific conductance is a physical measure but essentially reflects water chemistry based on the relative amount of ions dissolved in water.

Suspended Sediment

The physical quality of naturally occurring surface water is determined largely by the amount of sediment that it carries. As discussed in Chapter 9, the *total sediment load* in streamflow comprises suspended sediment and bedload. Of the two components, suspended sediment has a greater effect on water-column quality because it restricts sunlight from reaching photosynthetic plants measured as turbidity, whereas bedload has a greater effect on fish habitat. Sediment can also affect aquatic ecosystems adversely by smothering benthic communities and covering gravels that are often important spawning habitat for fish; this phenomenon is defined as *embeddedness*. Furthermore, sediment can carry undesirable concentrations of nutrients and heavy metals that affect water quality.

Sources

An increase in the sediment load of streams is one of the most widespread causes of degradation in the quality of water from forest, woodland, rangeland, urban, and agricultural watersheds. Surface water from undisturbed forest watersheds has relatively low suspended-sediment concentrations (generally <20 ppm). While much of this sediment originates in the stream channel itself, sediment can also be contributed from surface runoff during large storm events. Where organic soils are prevalent, much of the suspended material in streams is organic particles rather than mineral particles. Higher concentrations of suspended sediment are often the result of accelerated erosion caused by disturbances in drainage areas such as timber-harvesting operations, overgrazing by livestock, road construction, agricultural cultivation, urban expansion, or natural catastrophes including wildfires, large floods, and landslides.

Some of the more important ecological impacts of land-use activities involve physical changes in stream structure such as increased content of fine particles in gravel beds, erosion of streambanks, increases in stream width, decreases in stream depth, and fewer deep pools (MacDonald et al., 1991; Magner et al., 2008). Roads are the major contributors of sediment in streams; therefore, proper road design and maintenance are critical to minimizing sediment problems (Binkley and Brown, 1993). These problems are compounded when disturbances take place on steep terrain and near stream channels or lakes.

Nutrient and Heavy-Metal Transport Capacities of Sediment

Nutrient and heavy-metal losses from upland watersheds are usually measured by dissolved ion concentrations. However, a potentially important source of nutrient and heavy-metal loss and one that is often ignored is through its transportation by sediment. The losses occur as a result of the weathering forces of the physical and biological environment, the latter represented by the vegetation type on the watershed acting upon the parent bedrock. Pesticides such as atrazine are also known to adsorb to soil particles and can be transported by the water system in this manner.

Transported sediment from watersheds composed of different bedrock and vegetation combinations can carry high levels of nutrients and heavy metals. For example, a study of sediment from upland watersheds with limestone, granite, basalt, and sandstone geologies in the southwestern United States shows that, in general, limestone is high in *Ca* and *K*, while basalt is high in sodium (*Na*). *Mg* is highest in the sand fraction (0.061–2.0 mm) of basalt and the clay-silt fraction (<0.061 mm) of limestone (Gosz et al., 1980). Often, sandstone has the lowest concentration of these elements although high in silica (*Si*), and granite is frequently intermediate. The nutrients adsorbed to sediment particles can be indicative of the type of geologic formation in an area. Vegetation types on a watershed of a given geology primarily affect the organic matter content, total phosphorus (*P*), and levels of extractable nutrients of the sediments.

Variations in heavy-metal (zinc [*Zn*], iron [*Fe*], copper [*Cu*], manganese [*Mn*], lead [*Pb*], and cadmium [*Cd*]) levels in a stream are correlated with variations in sediment concentrations in many instances. In the southwestern United States, for example, sediment from different geologic strata has increasing heavy-metal concentrations in the following order: sandstone, granite, limestone, and basalt (Gosz et al., 1980). From the standpoint of watershed management, land-use practices that increase sediment production can increase nutrient and heavy-metal movement with suspended sediment.

Sediment transport of *P* can reduce the chemical quality of surface waters and, as a consequence, result in substantial changes in aquatic ecosystems. P is a limiting nutrient in many streams and lakes. When *P* loading increases, *eutrophication*, that is, the process of nutrient enrichment leading to dense algae growth can be accelerated. The resulting increase in algae and biomass in water systems can cause “dramatic adverse changes” in water quality. For example, small pine-covered watersheds in Mississippi yielded less than 60 mg/L of suspended sediment but the *P* concentration averaged from 329 to 515 μg/g of sediment (Duffy et al., 1986). These concentrations of *P* were from 2 to 3.5 times greater than the concentration found in the soils of the watersheds. Most of the sediment was transported in stormflow events when more than 70% of total *P* export and more than 40% of the total *N* was exported to illustrate the importance of maintaining low sediment yields with respect to the chemical quality of water.

Thermal Pollution

Water temperature can be a critical water-quality characteristic in many streams. The temperature of water, particularly temperature extremes, can control the survival of certain flora and fauna residing in a body of water. The type, quantity, and well-being of aquatic flora and fauna will frequently change with a change in water temperature. In several states in the United States distinctions between beneficial uses are made based on the temperature requirements of cold-water and warm-water fish. Of particular concern is a temperature increase due to land-use practices that could degrade a cold-water fishery. In general, an increase in water temperature causes an increase in the biological activity, which, in turn, places a greater demand on the *DO* in a stream. The solubility of oxygen in water is related inversely to temperature, which has a compounding effect (Table 11.1). Changes in water temperature can result in the replacement of existing species, such as cold-water trout being replaced by warm-water bass.

Clearing of riparian vegetation is one way in which water temperature can increase. The removal of trees along a streambank increases the exposure to solar radiation with the rise in water temperature predicted if one considers an energy budget for the water in the stream. If trees are removed from the streambanks, the only change in the energy budget is an increase in solar energy entering the system. This increase in solar energy will cause a

TABLE 11.1. Relationship between the saturated solubility of oxygen in water and water temperature

Water temperature (°C)	Solubility of O_2 (mg/L)
5	12.8
10	11.3
20	9.0
25	8.2

rise in water temperature because there are no new outlets of energy from the solar system. Increases in stream temperature can range from fractions of a degree centigrade for small openings in a forest overstory to more than 10°C for a complete removal of trees along the streambanks. Studies in the northeastern and northwestern United States have reported annual maximum stream temperatures to rise as much as 4°C and 15°C when riparian vegetation was removed from small streams.

Brown (1980) determined that the potential change in daily temperature due to streambank vegetation removal could be estimated from the following:

$$\Delta T = \frac{AR_n}{Q}0.000267 \tag{11.1}$$

where ΔT is the maximum potential daily temperature change due to exposure of a section of stream to direct solar radiation, in °F; A is the surface area of stream newly exposed to direct radiation (ft^2); Q is the streamflow discharge (ft^3/s); and R_n is the net solar radiation received by water surface that is newly exposed (BTU ft^2/min).

Mathematical models to predict stream temperatures, following modifications in the vegetative cover that shade streams, have been used on upland watersheds in the western United States. These models generally describe the physical situation of a stream including the vegetation bordering it. Changes in these variables as might occur through implementation of watershed management practices will often result in corresponding changes in water temperature. Variables in the models can be repeatedly changed to determine the possible effects of different methods of timber harvesting. Predicting water temperature changes can also be used to estimate corresponding changes in *DO* and subsequent impacts on aquatic flora and fauna. A more complete understanding of the water-quality consequences of changing riparian vegetation can be attained by doing so.

Dissolved Oxygen

The *DO* content in water has an effect on the aquatic organisms and chemical reactions that occur within the water body. The *DO* concentration of a water body is determined by the solubility of oxygen, which is inversely related to water temperature (Table 11.1), pressure, and biological activity. The solubility of oxygen in water can be estimated from the equation presented by Churchill et al. (1962):

$$O_s = 14.652 - 0.41022T + 0.0079910T^2 - 0.000077774T^3 \tag{11.2}$$

where O_s is the solubility of oxygen (mg/L) and T is the temperature of water (°C).

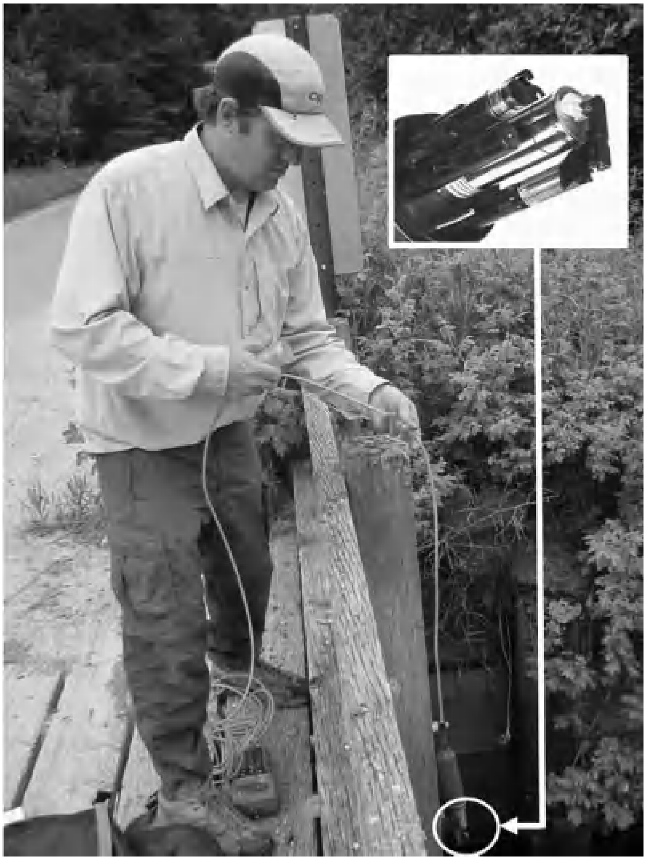

FIGURE 11.1. Measurement of *DO* with a multiparameter sonde (Photograph by Mark Davidson)

DO is a transient property that can fluctuate rapidly in time and space. From a biological perspective, it is one of the most important water-quality characteristics in the aquatic environment. *DO* concentration represents the status of the water system and is typically measured continuously using an *in situ* sonde which captures diurnal variation. However, point-in-time measurements are still made using the Winkler method to calibrate the *DO* probe. A *sonde* is a multiparameter device that can hold several probes and make simultaneous measurements of parameters. The sonde is a self-contained unit that can record and store data. Typically, a cable is connected between the sonde and a hand-held read-out device or directly to a data logger (Fig. 11.1). A commercially available sonde will include *pH*, specific conductance, and temperature; depending on the users' need the sonde will also include *DO*, salinity, turbidity or oxidation reduction potential (*ORP*).

The decomposition of organic debris in water is a slow process; therefore, the resulting changes in oxygen status respond slowly as well. Methods have been developed that estimate the demand or requirement of a given water body for oxygen. In essence, this is an indication of the pollutant load with respect to oxygen requirements and includes measurement of biochemical oxygen demand (*BOD*) or chemical oxygen demand (*COD*).

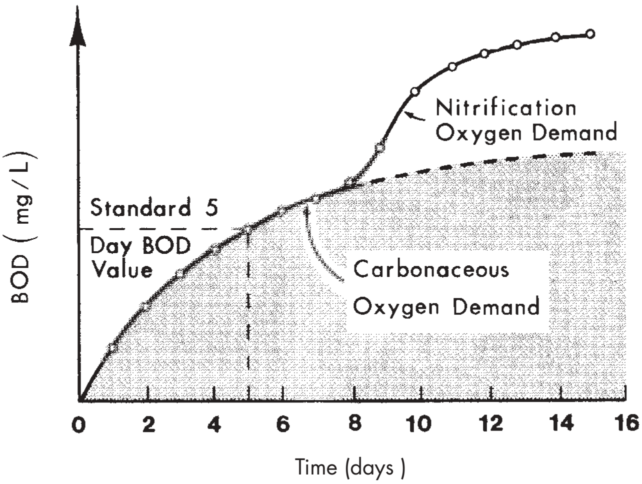

FIGURE 11.2. A *BOD* curve illustrating the carbonaceous demand phase and nitrification phase

Biochemical and Chemical Oxygen Demands

The *BOD* is an index of the oxygen-demanding properties of biodegradable material in water. Samples of water are taken from the stream and incubated in the laboratory at 20°C, after which the residual *DO* is measured. The *BOD* curve in Figure 11.2 illustrates the two-stage characteristic that is typical – the first stage is related to carbonaceous demand, and the second stage to nitrification. These two stages refer to the oxygen required to oxidize carbon compounds and nitrogen compounds, respectively. Unless specified otherwise, *BOD* values usually refer to the standard 5-day value, which is the carbonaceous stage. Such values are useful in assessing stream pollution loads and for comparison purposes (Table 11.2).

COD is a measure of the pollutant loading in terms of complete chemical oxidation using strong oxidizing agents. *COD* can be determined quickly because, unlike *BOD*, it

TABLE 11.2. Examples of biochemical oxygen demand (*BOD*) values for different conditions

Condition	*BOD* (mg/L) 5-day	*BOD* (mg/L) 90-day
Clean, undisturbed natural stream	<4	–
Effluent		
Pulp and paper processing	20–20,000	–
Feedlots	400–2000	–
Untreated sewage	100–400	–
Logging residue (needles, twigs, and leaves)	36–80	115–287

Source: Adapted from Ponce (1974), Dunne and Leopold (1978), and others.

TABLE 11.3. Examples of end products from organic loading in water bodies under aerobic and anaerobic conditions

	End products	
Types of compounds in organic load	**Aerobic**	**Anaerobic**
Carbonaceous (cellulose, sugars, etc.)	CO_2, energy, water	Organic acids, ethyl alcohol, methane (CH_4)
Nitrogenous (proteins, amino acids)	NO_3^{--} (in presence of bacteria)	NH_4^+, OH^- (NO_2^- temporary)
Sulfurous	SO_4^{--}	H_2S

does not rely on bacteriological action. However, *COD* is not necessarily a good index of the oxygen-demanding properties of materials in natural waters. Therefore, *BOD* is normally used for this purpose.

When organic material such as human sewage, livestock waste, organic-rich fine sediment or logging debris is added to a water body, bacteria and other organisms begin to break down that material to more stable chemical compounds. If oxygen is readily available and mixed in the water and the organic loading is not too great, oxidation can proceed without any detrimental reduction in *DO*. When oxygen is limiting or loading is too great, anaerobic processes can occur, which result in a less efficient oxidation process with undesirable byproducts in the water (Table 11.3).

A hypothetical sequence of changes that occur downstream of a heavy pollutant loading of biodegradable material is illustrated in Figures 11.3–11.5. These figures show schematically the effects of discharging raw domestic sewage, from a community of about 40,000 people, into a stream with a flow of 100 cfs (2.8 m^3/s). The *BOD* increases instantly at the point of discharge, which is followed downstream or in time with a reduction in *DO* concentrations (Fig. 11.3). The reduction in *DO* steepens the gradient of oxygen between the atmosphere and the water body, increasing the reaeration rate. The *DO* reduction curve is at a minimum in the area undergoing active decomposition. As reaeration takes place, *DO* concentrations increase and eventually reach *DO* levels before pollution.

Bacterial growth proceeds exponentially in the degradation and active decomposition zones. The corresponding decomposition of nitrogenous-organic matter takes place according to oxygen levels in the stream (Fig. 11.4). Other organisms respond to the modified environment particularly those organisms adapted to conditions of low oxygen, low levels of light, and high concentrations of organic material. Although many organism populations return to prepollution levels, some organisms do not. Algae populations can flourish because of the higher nutrient levels in the recovery and downstream clean-water zone. As a result, the habitat for higher organisms is modified to the extent that species diversity does not fully recover to the upstream, prepollution conditions (Fig. 11.5). Also, note that population levels of certain adapted species increase in the active decomposition and recovery zones because of limited competition. As water-quality conditions improve, the diversity of species recovers and populations of individuals within species drop to prepollution levels.

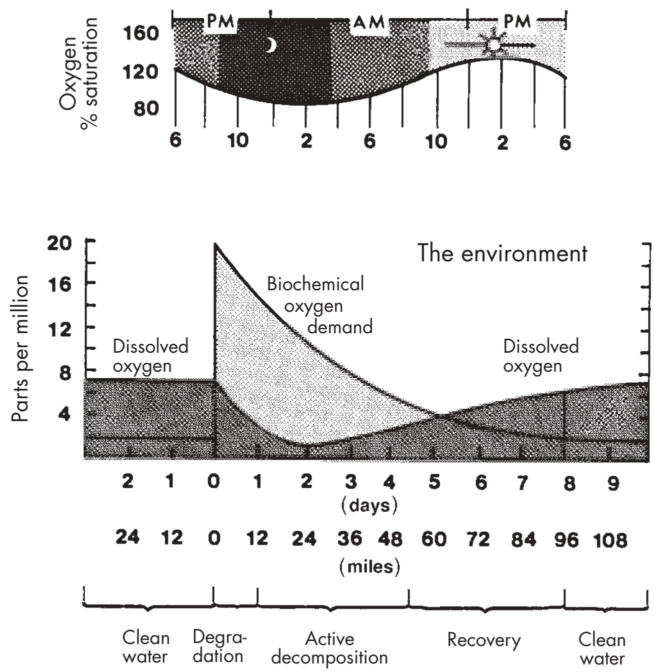

FIGURE 11.3. Effects of disposal of raw sewage in a stream on the *DO* and *BOD* for stream water, either in time or downstream. The effects of reaeration rate and the diurnal characteristics of *DO* are also shown (from Bartsch and Ingram, 1959, © Public Works J. Corp., with permission)

Field-Measured Indicators and Other Physical Characteristics of Water Quality

Several other characteristics including *pH*, acidity, alkalinity, specific conductance, and turbidity describe the physical condition of water. These characteristics can be important indicators of water quality and directly affect the chemical and biological condition of natural waters. These characteristics are fundamental parameters that are most often measured in the field and can be informative about chemical processes influencing water quality.

pH

The negative log (base 10) of the hydrogen-ion activity in moles per liter (in water) is defined and presented as *pH* and can be expressed as: $pH = -\log [H^+]$ or $\log (1/[H^+])$ of the H^+-ion concentration or activity. Typically, where a *pH* of $7.00 = -\log [1.0 \times 10^{-7}]$ or equal parts of $[H^+]$ and $[OH^-]$ (hydroxide); for example, a *pH* of $3.76 = -\log [1.7 \times 10^{-4}]$ of $[H^+]$ concentration which contains several orders of magnitude more H^+ than OH^- ions.

A *pH* value greater than 7 is indicative of *alkaline* water that normally occurs when carbonate or bicarbonate ions are present. A *pH* less than 7 such as 3.76 presented above defined as *acidic* water, indicates a significant amount of H^+ in solution compared to

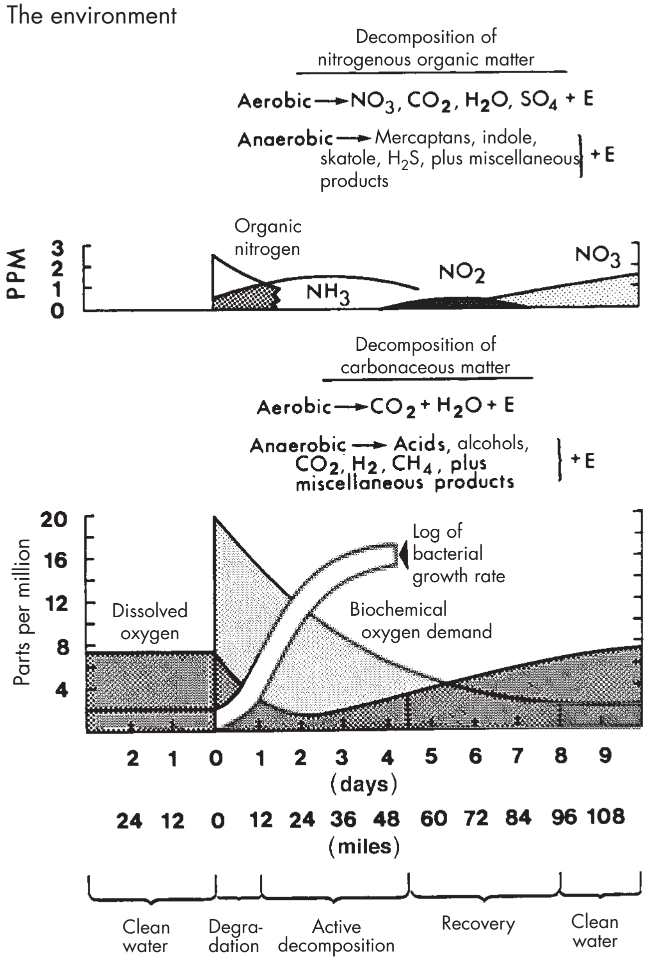

FIGURE 11.4. The relationship of accelerated bacterial growth to changes in *DO* and *BOD* due to disposal of raw sewage in a stream (from Bartsch and Ingram, 1959, © Public Works J. Corp., with permission)

hydroxide. When the pH of water affects and/or is affected by chemical reactions, aquatic systems respond depending on tolerance and thresholds for certain aquatic organisms; so it is important to examine some of the more common chemical relationships associated with pH. CO_2 reactions are some of the most important in establishing the pH level in natural waters. When CO_2 enters water either from the atmosphere or by respiration of plants, carbonic acid is formed, which dissociates into bicarbonate; carbonate and H^+ ions are then liberated to influence pH:

$$CO_2 + H_2O \leftrightarrow H_2CO_3 \leftrightarrow H^+ + HCO_3^- \leftrightarrow 2H^+ + CO_3^{--} \quad \textbf{(11.3)}$$

The pH at a point in time is an indication of the balance of chemical equilibria in water and affects the availability of certain chemicals or nutrients in water for uptake by plants. The pH of water also directly affects fish and other aquatic life. Generally, toxic limits are

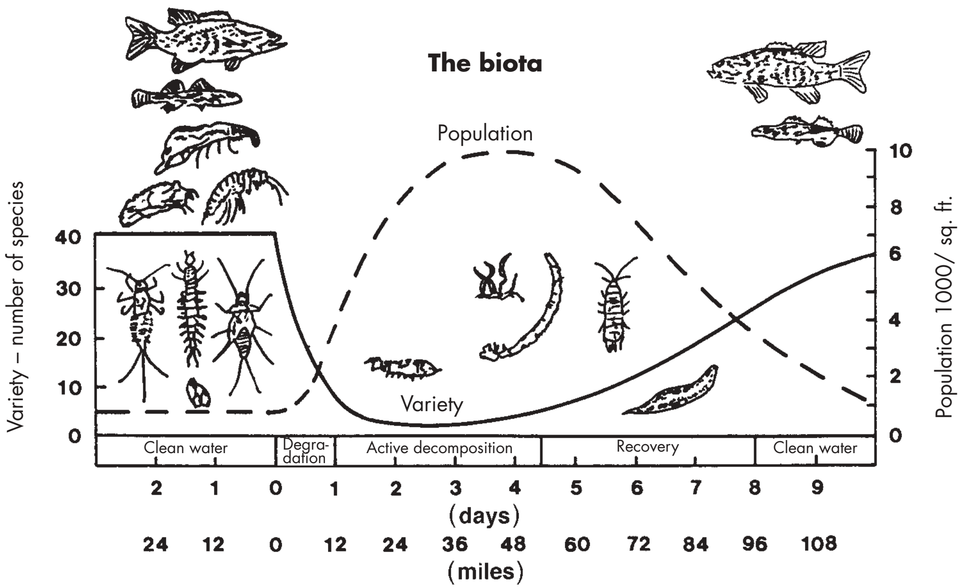

FIGURE 11.5. Effects of sewage disposal on species composition and populations of higher aquatic life-forms in a stream (from Bartsch and Ingram, 1959, © Public Works J. Corp., with permission)

pH values less than 4.8 and greater than 9.2. Most freshwater fish seem to tolerate *pH* values from 6.5 to 8.4; most algae cannot survive *pH* values greater than 8.5. A probe connected to a meter or sonde is used in the field to measure *pH*.

Acidity

Acidity and *pH* are closely related indicators of H^+ ion activity in water. Linked to *pH*, acidity is caused by the excess presence of free H^+ ions from carbonic, organic, sulfuric, nitric, and phosphoric acids. Acidity is important because it affects chemical and biological reactions and can contribute to the corrosiveness of water. Typically, we find acidity in rain, snow, bog water, and groundwater passing through granitic rock, some sandstones, and soils derived from non-calcareous minerals. This information is valuable when considering hydrologic pathways that influence water quality.

Alkalinity and Partial Pressure of CO_2 Gas

Alkalinity, the opposite of acidity, is the capacity of water to neutralize acid. Alkalinity is also linked to *pH* and is caused by the presence of carbonate, bicarbonate, and hydroxide that are formed when CO_2 is dissolved. A high alkalinity is associated with a high *pH* and excessive dissolved solids. When water is high in alkalinity, it is considered to be well buffered; that is, large amounts of acid are required to lower the *pH*. Another factor that influences alkalinity is *partial pressure of CO_2 gas (pCO_2)*. Typically in rainwater, the $pCO_2 = 10^{-3.5}$ (0.005) atmospheres of pressure. However, as rainwater passes through soil, the pCO_2 will increase to 10^{-2} (0.1) atmospheres of pressure. This process adds dissolved ions to the water.

As rainwater moves through the soil, the water attracts gaseous CO_2 from plant decomposition and Ca from calcite to form dissolved $CO_2 + H_2O = H + HCO_3$. This increase in bicarbonate requires the water to be charge balanced; meaning additional cations such as Ca must be drawn from the soil and/or rock. This increase in dissolved cations in the water drives the electrical conductivity (specific conductance) upward. Depending on minerals present in the soil or rock and the balance between deep infiltration and evapotranspiration the amount of bicarbonate alkalinity dissolved in water can vary. Most streams have alkalinities of less than 200 mg/L, although some groundwater aquifers can exceed 1000 mg/L when Ca, Mg, and Na concentrations are high. Ranges of alkalinity of 100–120 mg/L appear to support a well-diversified aquatic life. Alkalinity is best measured in the field by titrating with sulfuric acid to a pH of 4.5.

Specific Conductance

Specific conductance is the ability of water to conduct electrical current through a cube of water 1 cm on a side, expressed as micro mhos per cm at 25°C, or as micro siemens per cm (older instruments used the former units). By itself, this measure has little meaning in terms of water quality – except that specific conductance increases with dissolved solids as discussed earlier. Field measurement can reveal information about the past history of the water being measured. Water in a given region will typically have a normal range of specific conductance based on climate, geology, and land use. When values exceed the normal range, the data collected can suggest excessive cation or total dissolved solid (TDS) additions from pollutants such as ammonia or metals (Box 11.2).

Box 11.2

Water-Quality Influences from Row-Crop Agriculture in Southern Minnesota

Land-use changes over the last century in southern Minnesota have greatly influenced riverine water quality. Changes from tall-grass prairie to a mix of perennial and annual crops to the current corn/soybean base have resulted in ever more intensive land-use management. Along with drainage improvements, larger amounts of fertilizer have been applied to cropland to meet yield objectives. Both direct application of anhydrous ammonia, urea or animal manure adds large amounts of NH_4^+ to the soil. Large numbers of the cation NH_4^+ tends to displace other cations naturally present in the soil, for example, Ca and Mg. The ion-exchange process in the soil releases increasing ratios of Mg, Na, and Sr relative to Ca, the dominant cation (Magner and Alexander, 2002). Specific conductance typically runs 1.5–2.0 times higher under cropland compared to prairie soils. In time, the NH_4^+ is readily transformed in the soil by microbes and mineralization of soil organic matter releases relatively large amounts of NO_3-N. In perennial vegetation most of the NO_3-N is used by the plant material; however, under annual croplands 25–30% losses to leaching with transport to ditches and streams often occurs with soil thaw and early summer rains (Randall et al., 1997).

The objective of measuring specific conductance is the partial inference of the cation concentrations and causal candidates that might be present in the water. Specific conductance measurement is quick and inexpensive and can be used to estimate *TDS* as follows (Hem, 1992):

$$TDS\,(\text{ppm}) = A_o \times \text{specific conductance}\,(\text{micro mhos/cm}) \qquad \textbf{(11.4)}$$

where A_o is a conversion factor ranging from 0.55 to 0.75 with higher values associated with water high in SO_4 concentration.

Specific conductance is best measured in the field and typically is one of the parameters provided on a sonde (described earlier) along with *pH* and temperature. Specific conductivities in excess of 2000 micro Siemens/cm indicate a *TDS* level too high for most freshwater fish species. Values of seawater can exceed 50,000 micro siemens/cm.

Turbidity

The clarity of water is an indicator of water quality that relates to the ability of light to penetrate. *Turbidity* is an indicator of the property of water that causes light to become scattered or absorbed. The lower the turbidity, the deeper light can penetrate into a body of water and, hence, the greater the opportunity for photosynthesis and higher oxygen levels. Turbidity is caused by suspended clays, silts, organic matter, plankton, and other inorganic and organic particles. Instruments called *turbidimeters* are used to measure light penetration through a fixed water sample. Turbidity measurements, like specific conductance, can be used to estimate certain other factors affecting water quality. For example, the more easily measured turbidity can sometimes be used to predict suspended-sediment concentrations. A correlation between the two parameters is necessary, however.

Turbidity at one time was essential for making beer in Europe using water from the Danube River. The beer makers needed to know when the river water was clear enough to avoid undesirable additions to the composition of the beer such as clay particles. Turbidity is most useful for evaluating the suitability of human body contact with water such as swimming. Turbidity is best measured in the field and several vendors offer turbidity sensors as part of a sonde multiparameter probe package as described earlier. A simple alternative to a turbidity sensor is a secchi tube. A *secchi tube* is typically a 100 cm long plastic cylinder with a small secchi disk attached to a cord that the operator can move up and down to visually determine water clarity.

Dissolved Chemical Constituents

Natural streams are part of the areas that they drain and, therefore, are an integral part of a watershed and its ecosystems. Water is an effective solvent and as it comes into contact with each part of the system, the chemical characteristics of the water adjust accordingly. Chemical reactions and physical processes often occur simultaneously as the water contacts the atmosphere, soil, and biota. It is these reactions and processes and the condition of each compartment of the ecosystem that determine the kind and amount of chemical constituents in solution. For example, if the *pH* of precipitation changes, as previously discussed, the soil and organisms that live in the soil will adjust, which in turn, will alter the status of ion mobility within a watershed.

Streams that flow from undisturbed forested watersheds generally exhibit low concentrations of dissolved nutrients. Because of this, the biological productivity in most of these streams is also low compared to intensively managed cropland watersheds where fertilizer applications add elements like N, P, and K to the soil.

Sources of Nutrients

The major sources of dissolved chemical constituents in streamflow are the constituents in the precipitation falling on the watershed, geologic weathering of parent rock, and biological inputs. The cohesive properties of the bipolar-water molecule allow it to wet mineral surfaces and penetrate into the smallest of openings. Chemical and physical weathering then convert rock minerals into soluble or transportable forms that can be introduced into streams and lakes.

Biological inputs to water systems are primarily the photosynthetic production of organic materials from inorganic substances. Additional inputs are the breakdown of organic into inorganic compounds and the materials gathered elsewhere and subsequently deposited in the ecosystem by people and their animals. Leaf fall into streams is also an important source of organic matter that can result in periodic changes in nutrient concentrations. Some plants and particularly legumes can add nitrogen to the soil by fixing free-atmospheric N. Dissolved matter including organic compounds and mineral ions are added to the ecosystem by precipitation, dust, and other aerosols. Appreciable quantities of N, S^-, and other elements often occur in precipitation, dust, and dry fallout from the atmosphere between precipitation events. However, soils contribute the greatest amounts of dissolved chemical constituents to surface runoff and streamflow. As a result, land-use activities that affect soil properties and processes can affect the chemistry of streamflow.

In undisturbed ecosystems, the rock substrate and soil generally will control the relative concentrations of positive ions (cations such as Ca^{++}, Mg^{++}, K^+, and Na^+) in streamflow via subsurface discharge (see Chapter 7). Relationships between biological and biochemical processes in the soil govern the anionic (HCO_3^-, NO_3^-, and PO_4^{---}) yield. Anions such as chloride (Cl^-), nitrate (NO_3^-), and sulfate (SO_4^{--}), originate from the atmosphere and form minerals present in the soil, such as evaporite minerals (halite) or S^- minerals; pyrite or gypsum. Various geochemical and microbiological soil processes largely regulate ions in solution and the resulting composite mix that comprises both surface and subsurface flow chemistry.

Some of the more important dissolved chemicals or nutrients that occur in surface waters are described in the following paragraphs to aid in the understanding of the impacts of watershed management land use on water quality.

Nitrogen

Sources of N include the fixation of N gas by certain bacteria and plants, additions of organic matter to water bodies, and small amounts from the weathering of rocks. N occurs in several forms, including NH_4^+, gaseous NO, NO_2^-, and NO_3^-. Organic N breaks down into NH_4^+, which eventually becomes oxidized to NO_3^-, a form available to plants. In the absence of oxygen, the process of denitrification can convert NO_3^- back to NH_4^+ and N gas that typically escapes into the atmosphere. These processes are of considerable importance in intensively managed croplands where large amounts of N fertilizers are applied to croplands.

NO_3^- as nitrogen or NO_3-N is an ion of considerable interest in relation to land-use and watershed management practices. With the exception of P, most other ions remain

Table 11.4. **Mean concentrations of nitrogen and phosphorus for different land uses in the eastern United States**

Land use	Total *N*	Nitrate-*N*	Total *P*	Ortho-P
	mg/L			
Forest	0.95	0.23	0.014	0.006
Mostly forest	0.88	0.35	0.035	0.014
Mixed use	1.28	0.68	0.040	0.017
Mostly urban	1.29	1.25	0.066	0.033
Mostly agriculture	1.81	1.05	0.066	0.027
Agriculture	4.17	3.19	0.135	0.058

Source: From Omernik (1976).

at concentrations below water-quality standards (MacDonald et al., 1991). High concentrations of NO_3-N in water can stimulate growth of algae and other aquatic plants but if P is present, only about 0.30 mg/L of NO_3-N is needed for algal blooms. Some fish can be adversely affected when NO_3-N exceeds 4.2 mg/L. When NO_3^- levels exceed 45 mg/L (10 mg/L as NO_3-N) in drinking water, people's health can be affected – in particular, the blue baby disease. Drinking-water standards in the United States limit NO_3-N concentrations to 10 mg/L (Environmental Protection Agency, 1976). Some states and European countries have sought to find a more protective NO_3-N standard based on aquatic life and cancer risk.

Streamflow from undisturbed forest lands usually contain lower concentrations of N and NO_3-N than watersheds with other land uses (Table 11.4). Urban and agricultural development frequently increase NO_3-N and total N concentrations in streamflow. In particular, watersheds in the upper midwestern United States tend to have high levels of NO_3-N due to fertilizer additions to the soil for crop growth. High NO_3-N in midwestern streams and especially from small watersheds that are drained by subsurface pipes present a large challenge for watershed management and the management of gulf hypoxia (Magner et al., 2004). Gulf hypoxia is the eutrophication of gulf waters (e.g., Gulf of Mexico) by excessive nutrient runoff typically from cropland that causes low *DO*. The low *DO* due to *BOD* consumption limits aquatic life within the hypoxic zone.

Phosphorus

Phosphorus (P) originates from the weathering of igneous rocks, soil-leaching organic matter, and fertilizer inputs on intensively managed landscapes. Less is known about the P cycle than the N cycle, however. The common forms of P are essentially defined by the analytical technique used to quantify them rather than by natural processes. In an aquatic environment, P becomes available to plants by weathering and is taken up and converted into organic P. Upon decay, this process is reversed. However, along the way bio-available or soluble reactive P can fuel algae growth and reduce the transparency of lake water. P mobility is mostly linked with both organic and inorganic sediments and can be buried and temporarily removed from the P cycle. But, this does not mean that P remains removed from the cycle. P will move in saturated-subsurface flow under certain conditions, for example, chelation complex where P is bound by a negatively charged organic compound.

Oxidation–reduction processes can also influence the binding and release of *P* into an open water system, for example, bonding by *Fe* compounds.

P concentrations in streamflow are affected by land use just as *N* concentrations are (Table 11.4). Problems of eutrophication are often associated with accelerated loading of *P* to waters that are naturally deficient in *P*. Agricultural land use and urbanization result in the greatest problems of *P* loading to water bodies. Concentrations of *P* in wildland watersheds are typically low unless extremes in hydrologic events such as a drought followed by a flood flush *P* from normally stable wetland environments.

Calcium

Calcium (Ca^{++}) is one of the most abundant cations found in fresh waters because it is a major constituent of many rock types especially the carbonates such as limestone and dolomite. Exceptions are acid peat and swamp waters. As discussed earlier, soluble *Ca* is intrinsically linked to CO_2 present in soil water. *Ca* is one of the major ions contributing to hardness of water, *TDS*s, and specific conductance. High *Ca* concentrations do not appear to be harmful to fish and other aquatic life. Since *Ca* is typically present in most fresh-water environments, it can be used along with other cations and anions to infer the movement of water through a watershed (see Box 11.2).

Magnesium

Magnesium (Mg^{++}) similar to *Ca* is abundant in carbonate rocks such as dolomite. Its solubility increases with greater concentrations of CO_2 or a lower *pH*. The relationship between *Ca* and *Mg* can be helpful assessing water movement especially where open-water evaporation occurs within a watershed. *Ca*, for example, disassociates from $CaCO_3^-$ faster than *Mg* from minerals. This process will concentrate *Mg* in an open-water body such as a lake. The *Ca:Mg* ratio is approximately 4:1 or 3:1 in most carbonate-dominated geologic settings. In geologic settings where lake water has evaporated, the *Ca:Mg* ratio moves toward 1:1 or even 1:1.5. *Mg* can be toxic to some fish in concentrations greater than 100–400 mg/L. High concentrations of *Mg* along with other ions limit aquatic life in arid or historically arid geologic settings.

Sodium

Abundant in both igneous and sedimentary rocks, sodium (Na^+) is leached readily into surface and groundwater systems and remains in solution. *Na* is the cation most notably linked with common table salt (*NaCl*) that is also defined as the mineral halite. Dissolved *Na* in most fresh water is typically low, so higher levels of *Na* in surface or groundwater may be from human sources. Road salts are used to de-ice roads in colder climates and typically build-up in the soil adjacent to roads that can eventually enter both surface and groundwater. *Na* does not usually have any adverse impacts on fish unless both the *Na* concentration and the *K* concentration exceed 85 mg/L. Some levels of *Na* have beneficial effects by reducing the toxicity of aluminum and *K* salts to fish.

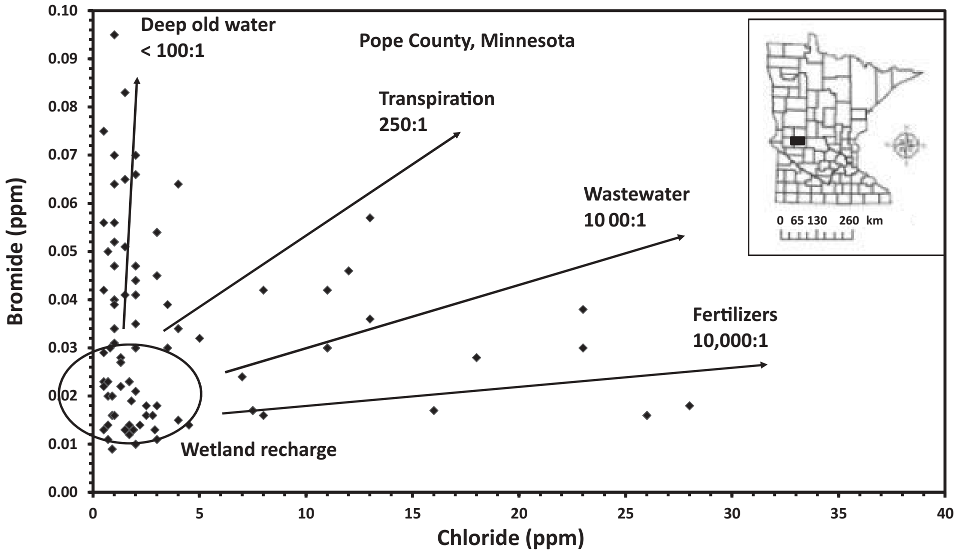

FIGURE 11.6. Concentrations of *Cl* are plotted against Br in Pope County, Minnesota (from Alexander, 2009)

Chloride

Chloride (Cl^-) is the most abundant of the halogens and combines with common metallic elements, such as alkali-earth metals that are readily soluble in water. Elevated concentrations of *Cl*, often associated with *Na* as described above, can be toxic to aquatic life. Therefore, it is of interest to be able to determine *Cl* concentrations in natural water bodies. Typically, bromide (Br^-) concentrations in most watershed waters remain stable with time and land-use change because *Br* concentrations are defined by historic geologic material and processes. Therefore comparing *Cl* to *Br* for a given water sample offers a tool for estimating the nature and type of *Cl* additions from human activity (Fig. 11.6).

The data in Figure 11.6 show that the water associated with upland wetlands is principally derived from precipitation and limited runoff and, therefore, has relatively low concentrations of both *Cl* and *Br*. Deep older groundwater that might be of glacial origin has natural *Cl:Br* concentrations of 100:1 or less. Transpiration processes increase the ratio to about 250:1; whereas, human additions, wastewater, and fertilizer greatly increase the ratio to 1000:1 and 10,000:1, respectively (from Alexander, 2009).

Potassium

Sources of potassium (K^+) include igneous rocks, clays, and glacial material. *K* is usually less abundant than *Na*, but it is essential to plant growth and is recycled by aquatic vegetation. Pristine waters generally contain less than 1.5 mg/L of *K*, but nutrient-enriched or eutrophic waters can contain more than 5 mg/L. Fish kill can occur when *K* levels exceed 400 mg/L, while levels greater than 700 mg/L can kill invertebrates as well. Since *K* is used in large amounts on intensively managed agricultural cropland, high concentrations can be found in these landscapes compared to wildland watersheds.

Manganese

Manganese (Mn^{++}), found in igneous rocks, is leached from the soil. *Mn* is essential to plant metabolism and is circulated organically to become soluble upon decay. At *pH* levels of 7 or less, the most dominant form is the divalent Mn^{++}. Concentrations of *Mn* rarely exceed 1 mg/L in undisturbed waters unless associated with highly reduced waters typically found in a wetland. The drinking-water standard for *Mn* is 0.05 mg/L. The measurement of *Mn* and Fe^{++}/Fe^{+++} in water is helpful for determining the relative oxidation state of a given sample of water. Elevated concentrations of *Mn* can suggest suboxic water. Suboxic water will not support fish because oxygen has been depleted. Field measurement of *DO* can infer suboxic condition, but the measurement of dissolved *Mn* and *Fe* using field colorimetric kits will provide a more definitive test of suboxic or reduced conditions.

Sulfur

Sulfur (S^-) occurs naturally in water from the leaching of gypsum and anhydrite found in glacial soil and other common igneous and sedimentary rocks. Weathering processes yield oxidized-sulfate (SO_4^{--}) ions that are soluble in water. SO_4^{--} is also found in rainfall at concentrations frequently exceeding 1 mg/L and sometimes greater than 10 mg/L. The higher concentrations of atmospheric SO_4^{--} are largely the result of air pollution and are the main contributors to acid precipitation.

Under reducing conditions, organic S^- can be converted to sulfide, for example,. pyrite. Metal sulfides occur with hydrogen sulfide (H_2S) being present below *pH* 7 and HS^- ions occurring in alkaline waters. H_2S produces a rotten-egg smell.

Water with desirable fish fauna generally contains less than 90 mg/L SO_4^{--}. Water with less than 0.5 mg/L will not support algal growth. Typical SO_4^{--} drinking-water standards are 250 mg/L.

Oxidation Reduction Potential

ORP refers to the potential for electron exchange to occur in biochemically mediated reactions. For the purpose of this book, reactions in water will only be discussed to illustrate the value of the metric. Since the presence of certain bacteria can influence the quality of water, estimating the *ORP* in water can guide land-use decisions for a hydrologist or watershed manager.

Electron equilibrium in water is typically microbiologically mediated to provide energy (food) for selected microbial communities. For example, *Fe*-related microbes require defined conditions to survive. If *Fe* is not present or too much oxygen is present, the microbes will not thrive. This occurs because the microbes live on electrons that become liberated when electron transfer occurs. The energy transfer can be illustrated by $2H^+ + 2e^-$ where the electrons are consumed as food by specific microbes poised in a microbe-specific environment. This phenomenon is similar to how people obtain energy from eating carbohydrates. One might ask what does this have to do with water quality. As stated earlier, low *DO* limits aquatic life and reduced water from a wetland can lower the *DO* concentration in a stream that would otherwise appear to support fish. For example, in the headwaters of the Mississippi River few fish can be found compared to locations downstream where more oxic water occurs.

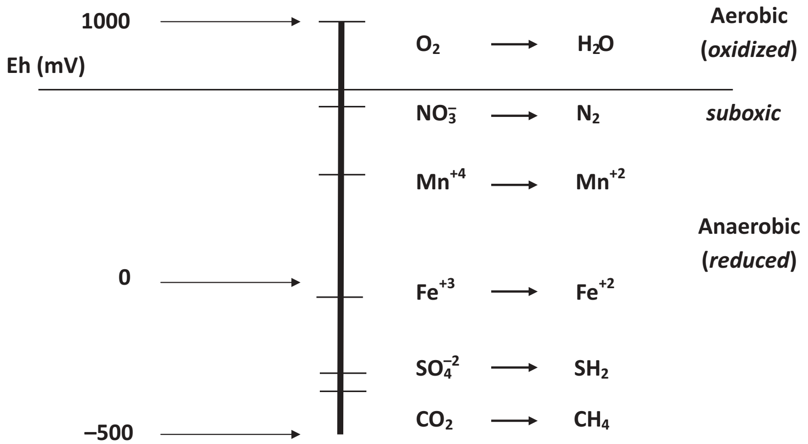

FIGURE 11.7. Aerobic conditions as measured by Eh in millivolts (mV), plot well above zero in the +300–+400 mV range. Suboxic waters plot above zero but below +300 mV and represent a transitional state between oxic and reduced waters. Anaerobic or reduced waters plot near and below zero and have very unique mV conditions associated with specific species (from Bouwer, 1994, as presented in Azadpour-Keeley et al., 1999)

One way to understand *ORP* is through the measurement of ions such as Fe^{++} and Fe^{+++}. Fe^{++} in a nonoxic water will transform to $Fe^{+++} + e^-$ (oxidation adds H^+ in solution) when water is exposed to the open oxidizing air where gaseous exchange occurs. In the field, historically a Standard Hydrogen Electrode would be used to measure *ORP*. However, these probes can become sluggish over time when placed into reduced waters – waters that contain little or no oxygen. A sluggish response occurs when the probe is then placed in oxic water; the probe reading can take 10 or more minutes to respond and provide a reliable value. Typically, the *ORP* probe never returns to its starting oxic-measuring resilience. Therefore, a probe rarely lasts more than a season of sampling.

Figure 11.7 illustrates the progression from oxic to suboxic to reduced conditions. In general, oxic, suboxic, and reduced conditions can be defined as follows:

- Oxic: +400 m*V*, high *DO* (rain water)
- Suboxic: <100 m*V*, low *DO* (<3–5 mg/L)
- Reduced: negative m*V*, no *DO*, creation of microbial created gases (wetland)

Pesticides and Fertilizers

Pesticides and fertilizers that are often used in agriculture and other types of land management have the potential to affect the quality of surface water and groundwater. These chemicals are introduced to accomplish management objectives by being applied to a specific *target* organism (pesticides) or location (fertilizers). Of concern to watershed managers are the chemicals that find their way into the water system and become transported to *nontarget* organisms. The risks or hazards to a nontarget organism are determined largely by the likelihood that the organism will come in contact with the chemical (exposure) and the toxicity of the chemical to that organism.

Table 11.5. **Values of 48-hour median tolerance (TL_M) for selected pesticides and aquatic organisms**

	Range of concentrations (*ppb*)		
Pesticide	**Aquatic insects**	**Crustacea**	**Fish**
Insecticide *DDT*	10–100	1–10	1–10
Endrin	0.1–1.0	10–100	0.1–10
Aldrin	1–10	10–100	1–100
Malathion	1–10	1–10	100–1000
Dieldrin	1–10	100–1000	1–100
Herbicide	1000–10,000	100–1000	1000–10,000
2,4-D (*BEE*)	–	–	100,000–1×10^6
2,4-D (amine salt) Picloram	10,000–100,000	10,000–100,000	10,000–100,000

Source: Adapted from Brown (1980) after Thut and Haydu (1971).

Toxicity effects can be either acute or chronic but are not necessarily lethal. *Acute* effects are those caused by exposure to large doses of a chemical over a short period, while *chronic* effects are those caused by exposure to relatively small doses of a chemical over a long period. One must realize that the characteristics of the chemical and the organism affected in addition to the size of the dose and the frequency and duration of contact all affect toxicity.

Toxicity to nontarget organisms is usually determined with bioassay techniques in which organisms are subjected to increasing concentrations of chemicals and observed over time. The concentration at which 50% of the organisms are killed is the lethal concentration (LC_{50}) or the median tolerance limit (TL_M). Values of TL_M for aquatic organisms and for commonly used chemicals are presented in Table 11.5.

Transport Processes

Dissolved chemical constituents can leave a terrestrial system by subsurface flow through the soil, surface runoff, or groundwater. These processes are part of nutrient cycling, which consists of inputs, outputs, and movement of dissolved solids and gases within the system. Outputs of nutrients in streamflow from upland watersheds that are transported downstream can cause water pollution for downstream uses. Therefore, land use and the various factors that influence nutrient cycling and the chemistry of water leaving upland watersheds are frequently a concern.

Movement of water through soil, along with the associated biological activity, controls the ionic composition of water leaving upland watersheds as streamflow. The most chemically active components in soil are clays and organic colloids. Clays have high exchange capacities in comparison to most other minerals in soil because of their large surface areas per unit of volume and their negative electrical charge. Organic colloids also have a large capacity to exchange ions in solutions for those adsorbed on their surfaces.

A simplified concept of the exchange processes of clays and organic colloids is that adsorbed cations are exchanged selectively for H^+ ions from the soil water. The H^+ ions, in turn, come principally from the solution of CO_2 in water and the dissociation of the resulting H_2CO_3 molecule into two H^+ ions and a bicarbonate ion. Dissolved CO_2 originates mainly from the metabolism of microorganisms and plant roots in the soil. Only water that remains for some time in the interstices among soil particles is likely to accumulate an appreciable load of dissolved CO_2 and, as a consequence, be effective in leaching mineral ions. It is the flushing of these accumulations that creates the initially high concentrations of cations during a stormflow event.

Ionic concentrations and their relationships to streamflow discharge are variable. For example, a positive relationship exists between H^+ and NO_3^- concentrations and streamflow discharge from forested watersheds in New England, but no relationship exists between discharge and Mg^{++}, Ca^{++}, SO_4^{--}, and K^+. When considered annually, individual cation concentrations appear to be independent of discharge rate on watersheds in the Appalachian Mountains of the eastern United States. Inverse relationships between ionic concentrations and discharge have been reported in selected streams of the Rocky Mountains and in California. Ionic concentrations in large streams are commonly lower at times of high discharge than at low discharge due in large part to the residence time of water in the soil. Dilution also occurs when large volumes of water move through soils or as overland runoff and direct precipitation on the stream. During periods of dry weather and low streamflow, the slow movement of water through soils enhances the opportunity for chemical reactions and soil-water concentrations of ions typically increase.

Biological Characteristics

The biological characteristics of water bodies are of interest from two perspectives: (1) the occurrence of pathogenic organisms that impact human health, and (2) the types and variety of biological organisms that can be used as indicators of the overall health of an aquatic ecosystem. Bacteria and protozoa in water are two types of the more important pathogenic organisms affecting human health. As described in the progression of changes illustrated in Figures 11.3–11.5, the effects of pollution on a water body can ultimately be related to the types, variety, and amount of aquatic biota that exist in that water body. Therefore, biota can be used as indicators of surface-water quality and the overall habitat characteristics of aquatic ecosystems.

Bacteria

Waterborne-pathogenic bacteria have long been recognized as causes of gastrointestinal illness in people, livestock, and wildlife species. Many of these diseases are transmitted only among members of the same species while others can be transmitted to hosts of different species. A major concern of watershed managers is that surface water from upland watersheds can become contaminated by health-impairing bacteria and pose hazards to the welfare of people and other animals.

Escherichia coli (*E. coli*) is a ubiquitous bacterium that is found in the gut of all warm-blooded animals. Long considered to be benign, several pathogenic strains of *E. coli* have evolved in recent years and gained attention due to the health risk that they pose. *E. coli* is often used as an indicator species in bacteriological testing to determine if water is suitable for drinking or other forms of human contact such as swimming. An estimate of their number, therefore, represents an index of bacteriological water quality. Other pathogenic organisms are difficult to trap, difficult to analyze and expensive to process (Buckhouse, 2000). As a consequence, testing for fecal coliform bacteria has been used as a surrogate sampling protocol with the understanding that if fecal coliforms are present, pathogens are then potentially present. This surrogate has led to false water-quality impairment listings and is no longer used by the State of Minnesota.

Knowledge of the cycles and variability of *E. coli* and other bacteria in natural waters and the relationships of bacteria to environmental factors on disturbed watershed lands are limited. However, investigations of bacteria-environmental relationships in relatively undisturbed areas have provided useful baseline information.

Protozoa

Waterborne-protozoan parasites can also cause gastrointestinal illness in people, their livestock, and wildlife species. *Giardia* and *Cryptosporidium* have drawn considerable attention (Buckhouse, 2000). Both parasites can be carried by a variety of warm-blooded animals including rodents, deer and elk, and livestock. One waterborne protozoan, *Cryptosporidium parvum* (*C. parvum*), has been implicated in large-scale outbreaks of gastroenteritis in humans (MacKenzie et al., 1994). Cattle are often perceived to be a leading source of *C. parvum* in surface water, although evidence supporting this claim is incomplete and contradictory in some cases (Atwill, 1996).

The possible presence of *Giardia* and *Cryptosporidium* in surface water has been known for decades but only recently has testing been conducted following gastrointestinal distress to determine whether these protozoa are the likely cause. However, watershed managers often have a difficult time determining if the *Cryptosporidium* found in samples of surface water is *C. parvum* or some other *Cryptosporidium* species not infectious to humans (Atwill, 1996).

Aquatic Biota as Indicators of Surface-Water Quality

Up to this point, we have discussed water-quality characteristics of a physical, chemical, and disease-causing nature. Any body of water can have a range of characteristics that are difficult to interpret. What has been sought in terms of water-quality management, therefore, is an integrative method of characterizing water quality; that is, a synthesis of the collective physical and chemical properties. Some investigators have suggested that the aquatic biota represent such a synthesis. The US Environmental Protection Agency (*EPA*) defines *biological integrity* as the ability of an aquatic ecosystem to support and maintain a balanced, integrated, and adaptive community of organisms having a species composition, diversity, and functional organization comparable to that of the natural habitat of a region (Karr and Dudley, 1981).

Any body of water in its habitat can support and maintain certain organisms. In general, the better the *aquatic health* of a water body as measured by an *Index of Biotic Integrity* (*IBI*), the greater the diversity of species of organisms (Karr and Dudley, 1981). Biologists have developed the *IBI* as a multiple-factor measure to assess the health of streams and wetlands in the last several decades. The factors that comprise the *IBI* are measures of the different components or functional attributes of the biological community that have been selected based on their ability to reflect change in their natural environment across a gradient of human disturbance.

As discussed with the earlier *BOD* loading example (Fig. 11.5), pollution can diminish a variety of organisms, leaving a habitat suitable for only a few well-adapted or “tolerant” organisms. As a result, a section of stream that supports a large variety of species of

Box 11.3

Indicating Water Quality by the Presence of Macroinvertebrates: An Index

Macroinvertebrate communities in streams have been found to be sensitive to stream disturbance and can be influenced by stream-channel morphology in the Pacific Northwest of the United States (Hershey and Lamberti, 1998). Using bioassessments of the abundance and diversity of macroinvertebrates in streams and lakes has been successfully used to monitor their condition or health throughout the world (Levy, 1998). The Hilsenhoff Biotic Index (*HBI*) is one example of a biological approach to monitoring water-quality conditions and biological potential that is based upon inventories of macroinvertebrates (Hilsenhoff, 1987). This index is based on the tolerance of a wide range of benthic macroinvertebrate species to a diversity of pollutants – some species are intolerant and receive a numerical score different from species that are tolerant.

organisms would be considered a better condition than one supporting a large number of a few species; that is, those adapted to the polluted environment. An exception to this concept is found in cold water where primarily cold-water species such as trout adapt to the thermal regime. The advantage of indicating surface-water quality through the presence of aquatic biota is that samples of chemical constituents, *DO*, temperature, and so forth only offer a snapshot picture at a point in time. The community of aquatic organisms represents more of a long-term indicator of overall impacts on water-quality and antecedent-flow conditions. It is this concept that underlies biological approaches to water-quality classification (Box 11.3).

The development of an *IBI* for a region must consider the following:

- The species used are commonly and widely distributed in all aquatic ecosystems (habitats) of the region.
- The species must respond with a range of sensitivities to varying types and levels of pollution. The *IBI* must exhibit differences across a gradient of human disturbance.
- The species complete their life cycles in streams, lakes, and wetlands. They must be continuously exposed to the physical, chemical, and biological conditions within their environment.
- Many organisms have well-known life histories. Therefore, *IBI* developers should understand how the biota will respond to different conditions that allow for the development of ranking schemes.

Using the above criteria the following list is an example of how an *IBI* could be developed from biological attributes discussed above.

- Taxonomic diversity – that is, biodiversity as discussed above.
- Number of intolerant groups – species that require a nonpolluted water condition to be present.

- Percentage of tolerant groups – species that can survive and even thrive in polluted waters.
- Percentage of dominant groups such as exotics, simple lithophilic spawners.
- Trophic structure – feeding behavior, that is, carnivores, omnivores, and generalists.
- Individual animal health – defined as *DELTs*, where D = deformities, E = eroded fins, L = lesions, T = tumors.

A partial list of *IBI* symptoms pointing to water-quality degradation is given below:

- Reduced populations of native species
- Fewer size (age) classes
- Reduced number of intolerant species
- Increased proportion of exotic species
- Reduced proportion of ecological specialists
- Simplified trophic web and interactions
- Increased incidence of serious disease and anomalies

How are these data used? The data are compiled and given a score that will separate the high from the moderate and low values assigned to each metric. The score from each metric is then added to provide an integrated total assessment score – this is the final number that represents the *IBI* for a given stream reach or other water body. The *IBI* numbers are then compared to other scores for similar regions based on regression analysis to assess the status and trends of riverine-surface waters. As stated earlier the underlying assumption is that the fish or macroinvertebrates will integrate biological, physical, and chemical stress upon the aquatic health of the system. Several states including Ohio, Maine, and Minnesota have developed regional indices of biological integrity (*IBIs*) using attributes of fish and macroinvertebrate communities. The states that have developed *IBIs* over the past decade or more use the information to conduct intensive watershed surveys and stressor identification follow up to support the *TMDL* process.

Stressor identification is a process (Cormier et al., 2000) of interpreting *IBI* data and discerning if human activity is causing a low *IBI* score. A low *IBI* score does not identify the stressor driving a poor biotic response as the stress might be due to intrinsic nature features in the regional geology and not land-use management. The *TMDL* process requires a comprehensive investigation to link biotic response with a stressor before a *TMDL* can be written. Furthermore, the *TMDL* must identify human stress that requires best management practice(s) to reduce the stress upon the aquatic community. The *TMDL* language is expressed in terms of load reduction (this is the *L* in *TMDL*), for example, reduce the load of fine-grained sediment filling interstitial pores of spawning gravel or reduce the load of *BOD* driving *DO* concentrations below biotic requirements for a given stream reach. A *TMDL* cannot be constructed based on a qualitative expression of impairment such as an *IBI* score; it must express a load reduction. Depending on the availability of data, each of the candidate causes are examined for causal pathways associated with stressors identified in the stressor identification process. The process must consider the following:

- Temporal co-occurrence of stress
- Spatial co-occurrence of stress
- Measured biological gradient of response
- Complete exposure pathway to a stressor
- Experiments from similar situations that can be cited from the literature

- A mechanism that links the stress to the organism
- Logical stressor-response plausibility
- Consistency of association
- Analogy
- Specificity of cause
- Predictive performance

The scientific team must develop a "prosecution case" using the criteria listed above and determine the strength of evidence to make the link between stress and *IBI* response. Some of the criteria might not be applicable depending on the lack of data. For example, pesticide data might not be considered because no pesticide was collected as part of the stressor identification study. A toxic chemical could be present in very low concentrations but a series of filtering tests would be required to justify an expensive and focused test for expensive organic compounds. Similar to a court trial that seeks to prove guilt or innocence, the stressor identification process must build a convincing argument to convict the primary stress/stressors. The primary stress/stressors must then be linked to a load reduction for the *TMDL*.

Groundwater Quality

The usefulness of groundwater for drinking or irrigation depends largely on its quality, which is largely related to the type and location of the aquifer. Water from igneous and metamorphic rocks is generally of excellent quality for drinking. Exceptions occur in arid areas where recharge water has high concentrations of salts because of high evaporation rates. The quality of water in sedimentary rocks varies. Deep marine deposits can yield saline water but shallow sandstone and carbonate rocks can have good-quality water.

Groundwater is generally of higher quality than surface water. Because it is in direct contact with rocks and soil material longer than most surface water, groundwater is usually higher in dissolved mineral salts such as Na, Ca, Mg, and K cations with anions of Cl, SO_4, and HCO_3. If such salts exceed 1000 ppm (mg/L), the water is considered *saline*. High concentrations of dissolved mineral salts can limit the use of groundwater for drinking because of its laxative effects.

Groundwater that contains high concentrations of Ca and Mg salts is defined as *hard water*. Its hardness is determined by the solubility of calcite or its equivalent as follows: soft water, 60 mg/L; moderately hard, 61–120 mg/L; hard, 121–180 mg/L; very hard, more than 180 mg/L. Although hard water leaves scaly mineral deposits inside pipes, boilers, and tanks and hampers washing because soap does not lather easily in hard water, it does not represent health hazards. Hard water is considered generally better for people's health than soft water.

Naturally occurring concentrations of Fe in groundwater can limit water use in some areas. Although not a health problem, Fe concentrations in excess of 0.3 mg/L affect the taste and color of water resulting in limited use for drinking, cooking, and washing clothes.

Unconsolidated deposits and other aquifers with high hydraulic conductivities such as limestone caverns and lava tubes can become contaminated from biological sources if they are close to surface sources of pollution (Box 11.4). People's garbage, sewage, and livestock wastes can contaminate such aquifers readily. Several factors limit our ability to solve groundwater contamination problems because of the variety of combinations

Box 11.4

Effects of Surface Land Use on Water Quality in Karst Aquifers
Many areas with Karst topography experience groundwater contamination from surface activities such as crop production and livestock grazing. Surface water and water occurring in limestone solution cavities are more directly connected than with most other aquifers; pollutants from the surface can move quickly into groundwater in such cases. Of particular concern in many agricultural and grazing areas in Karst topography is nitrate pollution. Boyer and Pasquarell (1995) reported the following levels of nitrate in Karst springs associated with different land use on the watersheds in the Appalachian region of the United States:

Average nitrate concentration (mg/L)	Grazing	Cropland	Forest	Urban/residential
	% of basin			
14	59	16	15	10
10	34	8	40	18
2.7	10	1	80	9
0.4	0	0	100	0

In Karst areas, best management practices that reduce surface water pollution can directly improve groundwater quality as well.

of land use, soil-water movement, biogeochemical reactions, and underlying geology. Furthermore, we often do not have reliable information on the rates and flow paths of groundwater. In addition, the location and extent of recharge areas are not well understood. Vastness, geologic complexity, and interconnectedness of most aquifers, coupled with these deficiencies, presents difficult and long-term challenges to protecting groundwater and cleaning polluted groundwater.

Cumulative Effects

Numerous activities on upland watersheds can affect the physical, chemical, and biological characteristics of water bodies. As discussed in this chapter many constituents can be considered separately as they affect water quality, but there can be considerable interaction and cumulative effects downstream that might not be readily apparent to watershed managers. A thorough discussion of cumulative watershed effects on water quality is beyond the scope of this book. Our intent is to discuss some examples upon which a student can begin to understand the complexity and importance of identifying cumulative effects.

The effects of environmental change on sediment have been considered in the previous chapters. However, it is also important to note that many chemicals and compounds are adsorbed and transported with silt or clay particles. As a consequence, land-use activities that increase erosion can also increase the transport of nutrients and chemicals from upland watersheds. The long-term productivity of these watershed lands can conceivably be reduced if these erosive processes continue.

The surface-water temperature at a stream site is determined by the temperature of inflowing water, radiant-energy inputs and outputs, conductive-energy inputs and outputs, and heat extraction by evaporation. Riparian (streambank) vegetation influences all of these heat exchanges in a stream (see Chapter 13). Beschta et al. (1987) and others have described relationships between land-use activities that reduce riparian vegetation and the resulting increases in stream temperatures. Water transports heat by changes in temperature from one site to another. Therefore, the effects of reduction in riparian vegetative covers on water temperature can impact downstream water uses that are dependent on temperature. Changes in water temperature also can affect *DO* and the habitat of various aquatic organisms – adding to any other effects on *DO* such as increased *BOD* loading.

Changes in land-use activities that reduce *DO* levels can also decrease habitability for fish. Increased nutrient loads deplete *DO* by stimulating algal blooms. Detrital-organic materials consume oxygen as they decay. The cumulative effects of all such factors on the oxygen and temperature regime of streams and rivers must be recognized throughout the watershed.

Because the transportation of dissolved chemicals in streams is largely dependent upon water movement, any land-use activity that alters the volumes or timing of runoff also affects the rates of chemical and nutrient transport. Timber harvesting often causes a short-term increase in nitrogen concentrations in streams (Mann et al., 1988; Tiedemann et al., 1988). The increases in nitrogen concentrations can result from reduced nutrient uptake, increased subsurface flow and increased alteration of *N* to leachable forms, increased volumes of decaying organisms, or combinations thereof. Baker et al. (1989) measured the rates at which nine nutrients were released from decomposing logging residues and found the most slowly released was *N*. Other management activities that contribute to these mechanisms such as herbicide application (Vitousek and Matson, 1985) and controlled burning (Sims et al., 1981) can have similar effects.

Some land-use activities alter existing chemicals to cause them to interact with the environment in different ways. For example, Helvey and Kochenderfer (1987) found that limestone-road gravels can moderate the *pH* of the runoff from acid rainfall. Birchall et al. (1989) discovered that aluminum released by acid rainfall is less toxic to fish populations if streams are enriched in *Si*. Bolton et al. (1990) found that vegetative conversions can alter the spatial distribution of the resulting *N* mineralization because the plants lend different mineralization potentials to the soil. Other points to consider are:

- Acid rainfall increases the mobility of nutrients and, therefore, contributes to nutrient deficiencies (Johnson et al., 1982).
- Chemicals introduced to stream channels can be redeposited downstream. Reservoirs can concentrate dissolved pollutants if evaporation rates are high.

SUMMARY AND LEARNING POINTS

The protection and maintenance of high-quality water are fundamental goals of watershed management. This chapter described water-quality characteristics and the impacts of land-use activities and disturbances on these characteristics and the aquatic ecosystem. After completing this chapter you should be able to:

- Define and explain the terms “water quality” and “pollution.”
- Discuss precipitation chemistry and its possible impacts on water quality.
- Identify and describe the different physical, chemical, and biological pollutants associated with different types of land use that can occur on upland watersheds.
- Describe how nutrients and heavy-metal concentrations and loadings to a stream can be accelerated above normal background levels. Discuss the implications.
- Describe how thermal pollution affects an aquatic system.
- Explain why *DO* is an important indicator of water quality. Discuss what happens when *DO* levels are substantially reduced, and define *BOD*.
- Describe what an *IBI* is, the stressor-identification process and how the information is used by government to write a *TMDL*.
- Describe how different actions and land use on a watershed can affect the physical, chemical, and biological characteristics of water that can cumulatively impact water quality.

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PART 3

Integrated Watershed Management

PHOTO 6. The Rodeo-Chediski fire in Arizona, USA, burned more than 189,000 ha of mostly ponderosa pine forest in 2002, resulting in serious soil and water impacts (Photograph by the US Forest Service) (For a color version of this photo, see the color plate section)

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

Managing upland watersheds located in forests, woodlands, and rangelands for sustaining or enhancing high-quality water yields is a primary objective of integrated watershed management (IWM). Included within these management practices can be sustaining or improving the hydrologic roles of riparian communities and wetlands. Recognition of the need to comply with established regulations and policies is paramount in planning and implementing IWM practices. Understanding the hydrologic changes that occur when wildland watersheds become fragmented by conversions to agricultural or urban land use is becoming increasingly important to IWM. Impacts of climatic variability and change on the management of watersheds must also be recognized. Furthermore, socioeconomic considerations rather than biophysical factors often determine whether IWM practices become implemented. Part 3 of this book considers these and other issues related to the success of achieving IWM.

The many facets of managing upland watersheds located in forests, woodlands, or rangelands are presented in Chapter 12. Included in this chapter are discussions on the effects of management on water yield, the streamflow regime, and water quality characteristics. The effects of fire, livestock grazing, and roads are also considered. Chapter 13 focuses on managing riparian communities and wetlands situated within watersheds. Basic riparian–watershed relationships and the hydrologic role of wetlands are considered in this chapter. The hydrologic effects of fragmenting watershed landscapes, the role of water harvesting to augment water supplies, the role of "Best Management Practices," a review of

PHOTO 7. Urbanization creates more surface runoff and compounds localized flooding – a 2011 scene of the Mississippi River at flood stage in St. Paul, Minnesota, USA (Photograph by Mark Davidson) (For a color version of this photo, see the color plate section)

regulations that can impact the selection of watershed management practices, and coping with climatic variability are among the topics discussed in Chapter 14. Socioeconomic considerations are discussed in Chapter 15 including policies and the policy process, a general planning process, and the framework for performing an economic assessment of watershed management practices. The tools and emerging technologies that are available for planning and implementing IWM practices are presented in the concluding chapter.

CHAPTER 12 *Managing Wildland Watersheds*

INTRODUCTION

We consider wildlands to include forest, woodland, and rangeland ecosystems that are not set aside for agriculture production, urban development, or large-scale mining operations. Wildland watersheds are managed to provide commodities such as wood, water, and forage and amenities including recreational opportunities, wildlife habitat, and aesthetic values. Varying combinations of these commodities and amenities are obtainable through integrated watershed management (IWM) practices. The hydrologic aspects of these management practices are the focus of this chapter. We will consider how watershed management practices, land use, and disturbances such as wildfire can change the amount and timing of streamflow discharges and the physical, chemical, and biological quality of water flowing from wildland watersheds.

FORESTS

Forests and the water flowing from forests are intrinsically linked. Precipitation falling on watershed landscapes is cycled through forest vegetation and soil to deliver streamflows that provide up to two-thirds of the water needed by the people of the USA (NRC, 2008). One of the reasons for establishment of the US Forest Service was to protect forests from degradation caused by people's activities or natural events and to sustain water flows. The quantity and quality of water flows from forests can be changed by timber-harvesting operations, silvicultural cuttings, the occurrence of wildfires, insect outbreaks, the incidence of disease, livestock grazing, urban development, and the construction of roads, trails, and other corridors. The US Forest Service maintains watershed research sites to monitor the long-term effects of vegetative modifications, land-use activities, climate change,

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

and fire on streamflow regimes, water quality, and other watershed values. Information about experimental forests and rangelands sites in the USA can be found in the website: http://www.fs.fed.us/research/efr/.

Forest Management and Water Yield

Forests yield the highest flows of water than other vegetative communities because they occur in areas that receive greater amounts of precipitation (Fig. 12.1). Therefore, forested watersheds possess greater potentials for increasing water yields than wildland watersheds supporting woodlands, scrublands, or grassland communities. When forest cover is altered as a result of management activities that clear or thin forest overstory, or from fires, insect infestations, and disease, changes in streamflow can occur.

Management of forests and development on forested watersheds often entails the construction and maintenance of roads, trails, and the rights-of-way for power and communication lines. Although the presence of these corridors can affect the hydrological functioning of the watershed, they would generally not be expected to significantly alter annual water yields (NRC, 2008) but could increase stormflow peaks (Megahan and King, 2004) as discussed later. This discussion, therefore, concentrates on water-yield response to forest-cover changes.

Studies conducted throughout the world have demonstrated that water yields (streamflow discharges) can increase when forest overstories on a watershed are cut or burned as summarized in Table 12.1. However, these streamflows generally return to their original levels once the watersheds recover from these impacts.

FIGURE 12.1. Streamflow from snowmelt runoff contributes much of the water originating on watersheds in the forests of the western region of the USA

TABLE 12.1. Increases in water yield associated with reductions in vegetative cover for noncloud forest or coastal forest conditions

Vegetative cover type	Increase in water yield per 10% reduction in cover		
	Average (mm)	Maximum (mm)	Minimum (mm)
Conifer and eucalypt[a]	40	65	20
Deciduous hardwood	25	40	6
Shrub	10	20	1

Source: Adapted from Bosch and Hewlett (1982), as reported by Gregersen et al. (1987).
[a] Pilgrim et al. (1982) indicated that in Australia pine used more water than eucalypts. Dunin and Mackay (1982) also indicated that interception losses of *Pinus radiata* were 10% more than eucalypt forests on an annual basis; in Australia annual differences between pine and eucalypt forests were estimated at 35–100 mm/year.

The increases in water-yield volumes associated with reductions in forest cover are caused mostly by reduced evapotranspiration (ET) losses. Up to 85–95% of the annual precipitation falling on forests can be intercepted and evaporated or transpired by trees, shrubs, and herbaceous plants, leaving 5–15% of the precipitation available for streamflow or groundwater recharge. ET losses are often reduced by changes in the structure, composition, or spatial arrangement of the tree overstory. Reduction in transpiration by removing tree or shrub overstory is determined by soil-water availability (Fig. 12.2). Changes in snowpack accumulation and melt patterns following these modifications can also increase water yields in cooler climates. Snowfall itself is not increased by these modifications but much of the snow that reaches the ground surface accumulates in snowpacks until conditions that are favorable for melting to occur. Increases in snowmelt runoff that flow to a stream channel increase perennial streamflows or initiate intermittent streamflows (Kattelmann and Ice, 2004). Up to 50% of the annual snowpack accumulation can be converted in streamflow originating from snowmelt-runoff events.

Water yields will often increase when (1) a tree overstory is removed by timber harvesting or thinning, (2) a tree overstory of comparatively high water-consuming trees is partially or completely replaced by lower water-consuming trees, shrubs or herbaceous plants, (3) deep-rooted tree species are replaced by shallow-rooted species of trees or shrubs, (4) tree species with high-interception capacities are replaced by species with lower-interception capacities, or (5) a combination of these actions occurs. Such changes can be due to planned forest management harvesting, or can occur from the effects of fires, insect infestations, or disease. The effects of forest management activities on water yields have been studied extensively in the temperate regions of the USA (Fig. 12.3) and elsewhere. These effects have been investigated to a lesser extent in tropical regions.

Timber-Harvesting Effects

Humid Temperate Regions. Water yield response to vegetation management in humid temperate climates would seem to be of questionable interest at first glance. There are situations, however, where either decreasing or increasing water yield can be of benefit. On one hand, areas with excessive water can call for management that does not increase water yield. On the other hand, increasing demands for water by expanding urban areas may warrant management of municipal watersheds for increased or sustained water yield, even

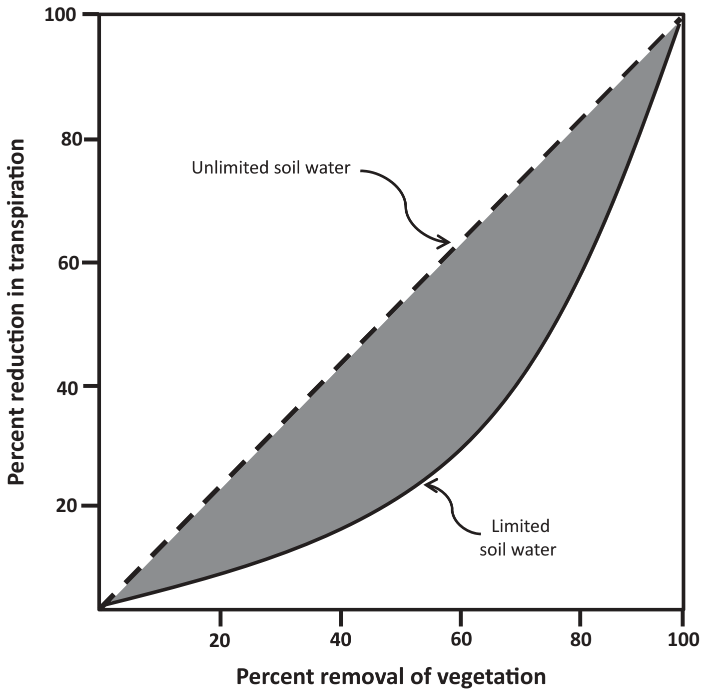

FIGURE 12.2. Hypothetical reduction in transpiration as a function of uniform removal (thinning) of trees and shrubs by thinning in conditions that vary from unlimited soil-water availability to plants (shaded area tending toward broken line) to limited soil-water availability (shaded portion tending toward solid line) as modified from Figure 4 in Hibbert (1979)

in areas with relatively high annual precipitation (>1000 mm). The following discusses relationships between vegetative cover and water yield in this climatic regime.

Studies in the hardwood forests in New Hampshire, USA, showed annual streamflow increases approaching 350 mm following the clearcutting of forest overstories in blocks, a pattern of strip cuttings, and complete clearcuttings of watersheds on the Hubbard Brook Experimental Forest (Hornbeck and Kochenderfer, 2004). Herbicides were applied to control the regrowth of trees in some of these studies but their effects cannot be separated from the tree-cutting activities. These increases in water yields are compromised when there was a change in the composition of the tree species or if hardwood species are converted to conifers.

An early study at the Coweeta Hydrologic Laboratory in North Carolina showed reductions of 20% in water yields because of the reductions in ET losses following the conversion of hardwood trees to pine on a watershed (Swank and Crossley, 1988; Jackson et al., 2004). Streamflow discharges after the conversion of a hardwood forest to grass on another watershed were similar to that of an original hardwood forest when production of grass on the converted watershed was comparatively high. The water yield then increased as grass productivity declined. Water yields increased 10% after all of the trees in a hardwood forest were cut but not removed from a third watershed without a road system. This finding

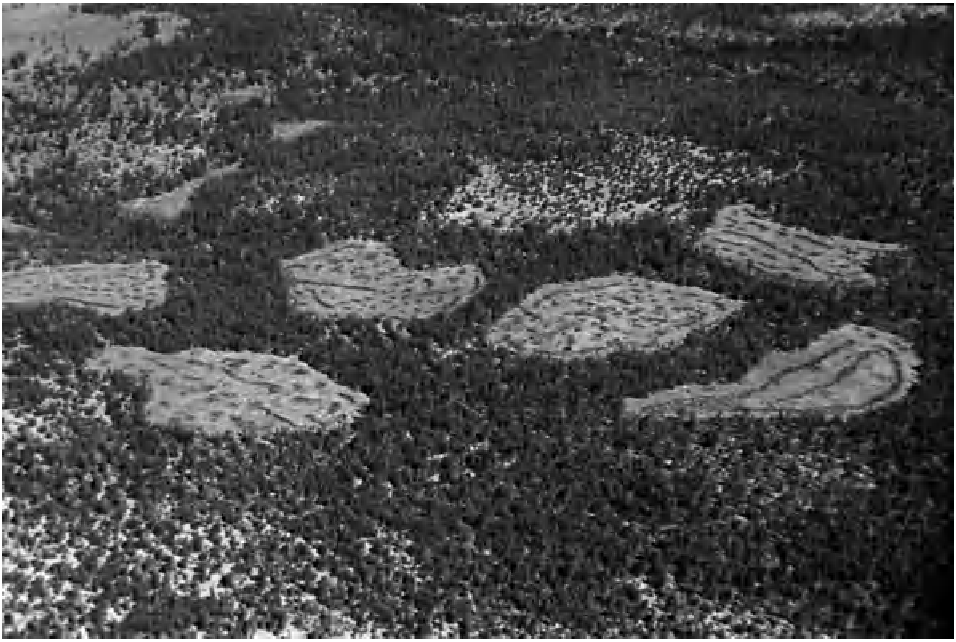

Figure 12.3. Clearcuts in ponderosa pine forests of Arizona, USA, have been shown to increase water yields

provided a baseline measure of the effects of cutting but not removing trees from a watershed on streamflow in the region.

Humid forest areas of the Pacific Northwest, USA, also show potential for augmenting water yield by manipulating forest cover. The climatic regime is different from that of the humid East. Annual precipitation can exceed 4000 mm at the higher elevations on the windward side of the Cascade Mountains. Much of this precipitation falls during the winter months and occurs as snow in the higher elevations. Distinctive dry periods usually occur during July through September.

Clearcutting Douglas-fir forests has increased annual water-yield 360–540 mm/year compared to 100–200 mm/year increases from partial clearcuts. Increases in yield diminish as forest vegetation grows back on the site. The water-yield increase expected for any year following clearcutting on one Douglas-fir watershed in Oregon was calculated as (Harr, 1983)

$$Y = 308.4 - 18.1(X_1) + 0.087(X_2) \tag{12.1}$$

where Y is the annual water-yield increase (mm); X_1 is the number of years after clearcutting; and X_2 is the annual precipitation (mm).

The above relationship would not apply for Douglas-fir forests in areas with persistent fog and long periods of low clouds; such situations exist in the coastal areas of the same region and show opposite responses to clearcutting. These "cloud forests" exhibit a reduction in water yield following clearcutting and will be discussed later.

The effects of timber harvesting and site preparation prior to conversion of native evergreen forests to plantations of Monterey pine on watersheds in high-rainfall areas of

New Zealand included "substantial" increases in water yields. When trees were clearcut on one watershed, a 550-mm increase in water yields was observed in the first year after the cutting (Rowe and Pearce, 1994). There was a 200- to 250-mm increase in water yields in the first year after the original tree overstory was clearcut and planted with Monterey pine seedlings in another watershed investigation. The rapid growth of bracken fern and honeysuckle following the tree cutting caused the increases in water yields to decline to pretreatment levels within a 5-year period in both of these studies.

The results of an investigation of the vegetation–water relationships at the Forest Hydrological Research Station in the Jonkershoek Valley of South Africa indicated an increase in water yields approaching 50% following a one-third reduction of the planted Monterey pine trees by a silvicultural thinning treatment (van der Zel, 1970). Most of the streamflow increase was attributed to the reduction in ET losses as a result of the thinning. This increase was sustained for only 3 years, however, at which time regrowth of the vegetation compromised the increase.

Forested Uplands in Snow-Dominated Regimes. Watersheds in mountainous regions exhibit a variety of soils, vegetation, and climate, which are accentuated by differences in elevation, slope, and aspect. As a rule, precipitation and water yield increase with elevation. Therefore, the greatest potential for increasing water yield usually lies in the mid- to upper-elevation watersheds. The proximity of many such watersheds to agricultural and urban centers in drier valleys downstream makes water-yield enhancement opportunities attractive to water resource managers. As a result, numerous watershed experiments have been conducted in the mountainous West (USA) to develop vegetative management schemes that increase water yield.

Opportunities for increasing water yield in the mountainous West depend largely on snow management as well as reducing ET. Much of the watershed research in this region has concentrated on timber-harvesting alternatives that redistribute the snowpack to achieve more runoff from snowmelt (Box 12.1). The reduced transpiration associated with timber harvesting increases runoff efficiency by leaving more water in the soil. Higher soil-water content during the fall and winter months results in a greater percentage of snowmelt ending up as streamflow rather than being stored in the soil.

Box 12.1

Effects of Strip-Cutting Forests on Snowpacks and Water Yield

Strip-cutting lodgepole pine forests in Colorado has been shown to affect wind patterns, snow accumulation, and melt (Gary, 1975). When strips were cut at a width equal to from one to five times the height of surrounding trees, snow-water equivalents were from 15% to 35% higher in the strips than in the adjacent forest. Nearly 30 years after harvest, average peak snow-water equivalent in the cut watershed still remained 9% above that of an uncut watershed (Troendle and King, 1985). Increases in annual flows above predicted levels are decreasing slowly in response to the regrowth of forests. Apparently, one-third of the increase in annual flows is attributed to this net increase in snow-water equivalent; two-thirds of the increase is largely due to reductions in ET.

Similar results have been observed in ponderosa pine forests in Arizona. The snow-water equivalent within the cut portion of a strip that was cut at a width equivalent to the height of the surrounding forest was increased 60%, or about 3 cm, compared to the uncut forest (Ffolliott and Thorud, 1974). In the figure below, the snowpack in the forest adjacent to the strip cut was reduced, resulting in a zone of influence that was much wider than the cut itself. Brown et al. (1974) reported on the cut of a mature stand of ponderosa pine on a watershed in the same area; strips equal to the height of the trees were cut in between strips two times the height of the trees. They found that this treatment increased water yield by about 3 cm and the effect was sustained for at least 5 years.

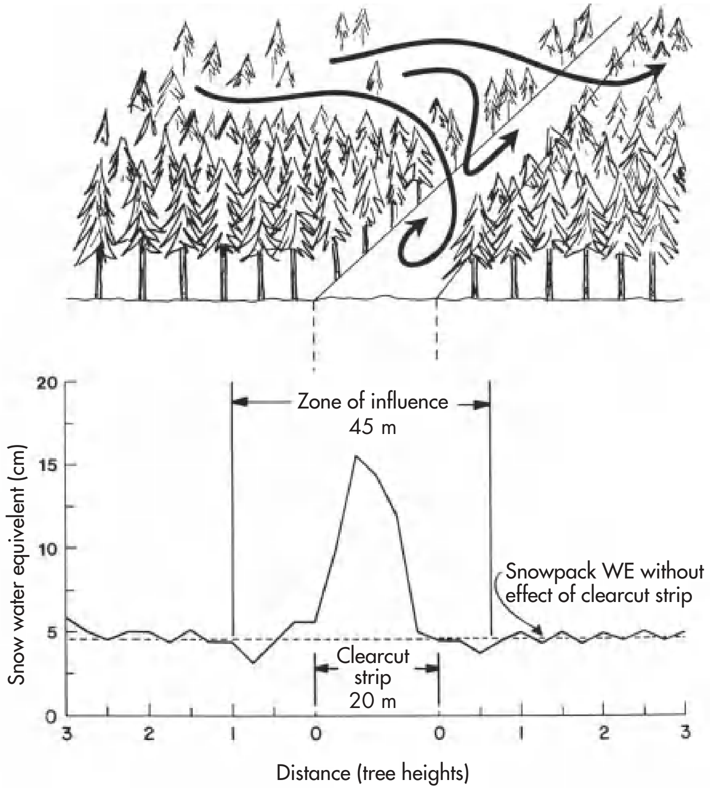

Figure. Effects of strip cutting a ponderosa pine forest in Arizona, USA, on wind patterns and snow deposition (adapted from Ffolliott and Thorud, 1974)

Streamflow from high-elevation watersheds in the Rocky Mountains exceeds 1000 mm/year. The region as a whole, however, yields less than 30 mm of streamflow per year with less than 15% of the land area contributing the majority of streamflow. These relationships prompted early research on the effects of changes in a high-elevation conifer forest on streamflow regimes by Bates and Henry (1928) in Colorado. This study was expanded to the Frasier Experimental Forest in the central Rocky Mountains to investigate the effects of modifying the subalpine forests on water yields. In one study of note, the findings from earlier plot studies at Frasier were extrapolated to a watershed-basis by cutting 40% of the forest on a 289 ha watershed in strips with intervening uncut strips (Gottfried et al., in process). The average water-yield increase of 80 mm or 40% that was observed in the first 28 years following the strip cutting was attributed to an increase in snowpack water equivalents because of the reduced interception loss and sublimation of snow. Although the increase in streamflow discharges declined later, it averaged 58 mm or 30% for 50 years after the treatment. A redistribution of the snowpack was again the assumed reason for this continuing increase.

In the Sierra Nevada range of the western USA, annual water yields vary from 350 to 1000 mm/year with the higher-elevation watersheds exhibiting the highest yields. If large forested watersheds were managed exclusively for water-yield improvement, annual harvesting schedules could increase water yields from 2% to 6% (Kattelmann et al., 1983). However, under multiple-use and sustained-yield harvesting schedules, annual water-yield increases of less than 20 mm would be expected.

The potential for increasing water yield from ponderosa pine forests in the southwestern USA has been of interest because of the scarcity of water and a rapidly expanding population. Depending upon the percentage of forest cover removed and annual precipitation, water-yield increases of 25–165 mm/year have been reported (Hibbert, 1983; Baker, 1999). The effects normally persist for only 3–7 years because of vegetative regrowth.

Forest harvesting effects on water yield are minimal in drier and generally lower elevations of the western USA. Unfortunately, these are the areas where water is usually scarce. An analysis by Hibbert (1983) showed that forest and other vegetative manipulations could increase water yields only on watersheds receiving >450 mm of annual precipitation. He reasoned that precipitation below this amount is effectively consumed by residual overstory vegetation and subsequent increases in herbaceous cover on the watersheds.

Tropical Regions. The effects of changing forest cover on water yield in the humid tropics are less well known than in temperate regions. This is surprising because of the interest in tropical forests and the influence that humid tropical forests exert on climatic conditions. However, long-term studies to better understand, appreciate, and quantify such inferences are limited. As a consequence, questions dealing with regional or global implications of vegetative changes in humid tropical forests have often been addressed more with conjecture than fact.

The few watershed-level experiments in the humid tropics have shown that the responses of water yields to changes in tree overstories are largely similar to those observed in temperate regions. For example, rainforests that have been logged and then cleared for pastureland in North Queensland, Australia, resulted in a 293 mm or 10% increase in water yields for more than 2 years after these interventions (Gilmour et al., 1982). Weekly comparisons of these changes indicated that minimum streamflow discharges increased from 15% to 60% following the logging and clearing activities. These streamflow responses were similar in magnitude to cleared forests in the temperate regions in the wet season.

The effects of forest removals on water yields in the humid tropics are generally shorter in duration than in temperate regions because of the rapid regrowth of vegetation. Therefore, forest removal might not be expected to affect water yields for more than a few years. Longer-term and substantial increases in water yields can result when forests are converted to agricultural croplands or pastures; however, with the larger the percentage of a watershed affected, the greater the increase in water yields.

Cloud Forests. Forests that occur along foggy coastal areas or those situated in montane fog or cloud belts near coastal regions in the tropics have unique hydrologic characteristics. In these areas of frequent and persistent low clouds or wind-blown fog, atmospheric moisture condenses on foliage and stems of trees where it coalesces into larger droplets that drip from the canopy to the forest floor. In addition to canopy drip, some of the droplets evaporate into the atmosphere and some flow down stems. These *cloud forests* contribute moisture in excess of the amount of annual rainfall, generally ranging from 5% to 20% of average rainfall but in some cases exceeding 1000 mm/year (Bruijnzeel, 2004).

Montane-cloud forests in Central America, which occur mostly on old volcanoes or high mountains, add significant amounts of water to local watersheds (Zadroga, 1981). Cloud forests that occur in narrow bands along the Pacific Ocean in northern Chile, though limited in extent, can contribute significantly to the annual water budget in this arid region. For example, in a region where the annual input of water is approximately 200 mm, measurements indicate that two-thirds of this amount is due to the interception of fog that flows inward from the Pacific Ocean and coalesces on the foliage of the trees. To increase this input further, artificial barriers have been constructed to intercept the movement of fog.

Streamflow increases that are often anticipated after forest removal can be limited and even decrease following the cutting of trees in coastal-cloud forests (Azevedo and Morgan, 1974; Lovett et al., 1982). On the Bull Run municipal watershed near Portland, Oregon, USA, for example, timber harvesting has reduced fog drip from trees on the watershed that resulted in reduced levels of dry-season streamflow (Harr, 1982). Fog drip from mature Douglas-fir forests has been observed to add nearly 880 mm of water each year (Harr, 1983). Isotopic studies in California and northern Kenya indicate that fog drip from trees of cloud forests contributes significantly to the recharge of groundwater in fog-laden areas (Ingraham and Matthews, 1995). Fog drip represented 35% of the total input of water to the coastal redwood forests of northern California over a 3-year study (Dawson, 1998).

From research in cloud forests we conclude that maintaining a forest overstory can be important in sustaining streamflow, groundwater recharge, and baseflow conditions of streams and rivers. Conversely, removal of forest overstory can diminish these flows.

Effects of Fire

The occurrence of fire throughout history has led to a body of knowledge on the effects of fire on hydrologic processes. Depending on its intensity, severity, and extent, a fire can change vegetative structures, modify soil properties, and impact components of the hydrologic cycle. Lightning-caused wildfire requiring a suppression response can result in a modification or replacement of the prefire forest cover and, in doing so, changes the hydrologic functioning of a watershed. Prescribed burning can also impact the hydrologic processes of a watershed but to a lesser extent than wildfire because of the low-fire severities involved.

The loss of the vegetative cover and organic-matter accumulations on the soil surface and the possible formation of water-repellent soils (DeBano, 1981) are among the main causative mechanisms for the often-observed increases in postfire streamflow discharges

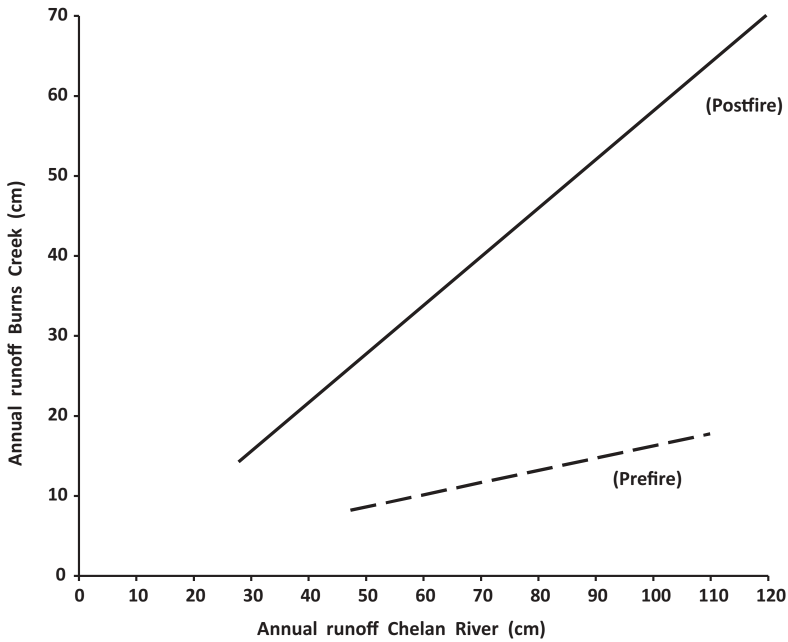

FIGURE 12.4. Annual streamflow discharges from a burned watershed in the Cascade Range of eastern Washington, USA, before and after a wildfire in relation to annual streamflow discharge from a control (adapted from Helvey, 1980)

(see Chapter 5). These effects contribute more to the generation of higher surface runoff, increased stormflow peaks and volumes, and sedimentation as discussed later – than to annual water yields. Increases in water yield from burned areas would generally be temporary and would be expected to return to preburn levels once forests have regenerated. Increases in water yields following fires would be generally greater in forests with inherently high ET losses.

Annual water yields from a 565-ha watershed in the Cascade Range of eastern Washington, USA, on which a wildfire destroyed almost 100% of the mixed conifer forest increased "dramatically" relative to a prefire relationship between streamflow from the watershed that was burned and that from a nearby unburned watershed (Helvey, 1980). Differences between the observed- and predicted-streamflow discharges varied from 105 mm in a dry year to 475 mm in a wet year (Fig. 12.4). Soil-water storage remained largely because of the abnormally high magnitudes of precipitation falling throughout the study with the burned and unburned watersheds becoming more sensitive to precipitation amounts as a consequence.

Following a wildfire of varying severities in a ponderosa pine forest of northern Arizona, Campbell et al. (1977) instrumented three small watersheds to evaluate the effects of the wildfire on water yields. Most of the trees on a severely burned watershed of 8.1 ha were destroyed, the wildfire was confined mostly to the litter and duff layers on a moderately burned watershed of 4 ha, and an 18.2-ha unburned watershed was the control in this evaluation. The annual streamflow discharge from the severely and moderately burned watersheds averaged 28 and 20 mm, respectively, compared to 5 mm from the control watershed in the 3 years of postfire monitoring. Nearly all of these streamflows resulted from seasonal snowmelt runoff in the spring.

Scott (1993) reported that the water yields from three watersheds in the Cape Region of South Africa increased by an average of 120 mm or 10% following a wildfire that consumed most of the fynbos shrubs and glasslike plants on these watersheds. The increase in streamflow was attributed to the reductions in interception and ET losses as a consequence of the wildfire. Increases of more than 70 mm or 10% in water yields and almost 6.5 mm or 60% in stormflow volumes were observed after a wildfire on another watershed in the Cape Region that had been converted from fynbos shrubs to a Monterey pine plantation before the burn (Scott, 1993). There were no changes in water yields after a wildfire swept across a third watershed in the region (Scott and Van Wyk, 1990). The low rainstorms that followed the burn limited postfire interpretations of the statistically weak calibration of stormflows with an unburned watershed.

There was a 60 mm or 30% increase in water yields in the first year after a wildfire burned a 145 ha watershed supporting maquis, cork oak, and chestnut trees in southern France near the Mediterranean Sea (Lavabre et al., 1993). The postfire increase in water yields was related mostly to the reduction in ET losses due to destruction of the tree cover.

In contrast to wildfires, prescribed burns are generally low-severity fires that normally do not destroy enough vegetation or consume enough litter and other decomposed organic matter to significantly alter water yields. For example, annual water yields did not to change significantly in comparison to prefire levels in the 6 years following a prescribed fire on about 45% of a 470-ha watershed in eastern Arizona supporting a ponderosa pine forest (Gottfried and DeBano, 1990). Analysis of the postfire streamflow measurements showed an increase of almost 9 mm for the entire watershed. If this increase was prorated to the burned area only, it would be equivalent to a water yield increase of almost 20 mm or about 20%. It cannot be assumed, however, that a more complete burn of the watershed would have resulted in significant changes in water yields. Similarly, controlled burning of understory vegetation and logging residues following timber harvesting to reduce the accumulations of flammable fuels would not be expected to have significant effects on annual water yield.

Effects of Insects and Disease

Outbreaks and incidences of insects and disease that alter forest biomass or cover conditions can affect streamflow. Mountain pine beetle infestations have damaged vast areas of coniferous forests in western North America, destroying more than 23 million ha of forests across the interior west of the USA and Canada (http://www.reuters.com/article/2011/01/23/us-beetles-forests-idUSTRE70M28S20110123). Such infestations have been promoted by warmer winter temperatures and when they occur in older forests that are weakened by drought, the trees become more susceptible to wildfires.

Bark beetle infestations that began in 1939 and continued through 1965 in watersheds of western Colorado killed 4-billion board feet of standing timber and increased streamflows in the White and Yampa rivers by an average of 31.2 to 23.6 mm/year from 1941 to 1965, respectively (Bethlahmy, 1974). The streamflow increases caused by such beetle outbreaks would likely be increased further if followed by wildfires raising the risk of excessive surface runoff, higher stormflows, and increased sediment transport to downstream areas. Conversely, wildfires can leave vast areas of weakened or dead trees that are susceptible to infestations by bark beetles. While some of these effects might be extrapolated from general relationships of forest dynamics and water (Uunila et al., 2006), there remains much to be learned about predicting the hydrological impacts of insects and diseases in forest ecosystems.

Estimating Changes in Water Yields

Changes in water yields following planned or unplanned modifications of a forest overstory can often be estimated by regional relationships, a water-budget approach, or applications of hydrologic computer-simulation models.

Regional Relationships. Regional relationships can be useful in approximating changes in water yields from watersheds where the climate, vegetation, soils, and topography are similar to the watershed where the relationships were developed originally. For example, Douglas (1983) developed regional relationships to estimate the responses of water yields to the cutting of trees in the hardwood forests and coniferous forests in the eastern USA. The response to cutting hardwood forests on watersheds is estimated as

$$Y_{\mathrm{H}} = 0.00224\left(\frac{BA}{PI}\right)^{1.4462} \quad \textbf{(12.2)}$$

$$D_{\mathrm{H}} = 1.57Y_{\mathrm{H}_1} \quad \textbf{(12.3)}$$

$$Y_{\mathrm{Hi}} = Y_{\mathrm{H}} + b\log(\mathrm{i}) \quad \textbf{(12.4)}$$

where Y_{H} is the first year increase in water yields (in) after cutting the hardwood forest (H); BA is the percent basal area removed cutting; PI is the annual potential solar radiation in (cal/cm^2 $\times$ 10^{-6}) for the watershed; D_{H} is the duration of the increase in water yields (yr); $Y_{\mathrm{H}i}$ is the increase in water yields for the ith year after the cutting (in); and b is the coefficient derived by solving Equation 12.3 when $I = D_{\mathrm{H}}$ and $Y_{\mathrm{Hi}} = 0$.

The relationships for coniferous forests are

$$Y_{\mathrm{C}} = Y_{\mathrm{H}} + (I_{\mathrm{C}} - I_{\mathrm{H}}) \quad \textbf{(12.5)}$$

$$D_{\mathrm{C}} = 12 \quad \textbf{(12.6)}$$

$$Y_{\mathrm{Ci}} = Y_{\mathrm{C}} + b\log(i) \quad \textbf{(12.7)}$$

where $I_{\mathrm{C}} - I_{\mathrm{H}}$ is the difference in interception between conifers and hardwoods (in).

These regional relationships apply for conditions where annual precipitation exceeds 1015 mm, the precipitation is distributed uniformly on the watershed, and the solar radiation indices (Lee, 1963) are between 0.20 and 0.34 that are the coefficients that correct for latitude, slope, and aspect.

Interest in converting aspen to conifer forests in the northern Lake States, USA, prompted questions about the effects of such conversions on water flow from the affected areas (Verry, 1976). Differences in interception between aspen and red pine were used to predict changes in net precipitation (water yield) using annual precipitation data and the basal area of aspen (*Populus* spp.) and red pine (*Pinus resinosa*) stands, respectively, as illustrated in Figure 12.5. For example, if annual precipitation at a site were 756 mm, the

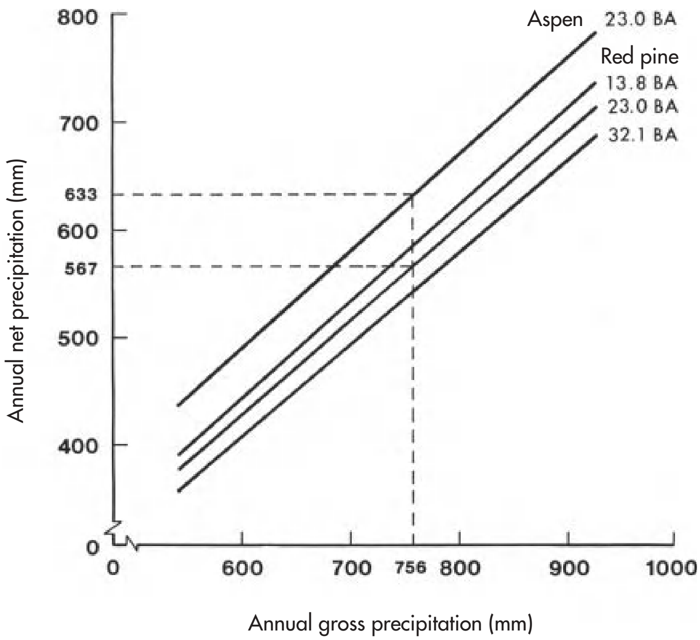

FIGURE 12.5. Relationship between net and gross annual precipitation for aspen and red pine (from Verry, 1976)

conversion of an aspen stand with a basal area of 23 m^2/ha to a red pine stand of the same basal area would reduce annual net precipitation by 66 mm (633–567 mm).

A Water-Budget Approach. A water budget can be used to estimate water yields from forested watersheds with comparatively high infiltration capacities and deep soils. Water yields from these watersheds are governed primarily by soil-moisture storage characteristics. Changes in water yields associated with changes in vegetative cover are estimated by a water-budget analysis by knowing the changes in effective rooting depths (Box 12.2).

The water-budget approach to estimating water yields can be modified when more information becomes available. For example, the temporal resolution of the estimation can be reduced from monthly to daily accounting. If seasonal *ET* relationships are known, the *ET–PET* relationship can be modified accordingly. Edwards and Blackie (1981) reported *ET*/*PET* ratios for different plant-soil systems in East Africa. These ratios could be used in that region to improve estimates of *ET* changes. Likewise, functions such as $ET = (PET)f(AW/AWC)$ as described in Chapter 4 can be used. The *AW* and *AWC* terms are related to the effective rooting depth of the respective vegetation types. Knowledge of transpiration response to soil-moisture conditions is assumed.

Applications of Hydrologic Computer-Simulation Models. Hydrologic computer-simulation models that can be used to estimate water yields from forested watersheds range from simplified-empirical relationships at one extreme to detailed process-oriented relationships at the other extreme. These simulation models often

Box 12.2

Application of a Water-Budget Method to Estimate Changes in Water Yield due to Clearcutting a Mature Hardwood Deciduous Forest

A clearcut of 190 ha of mixed hardwoods is to be considered on a watershed that drains into a water supply reservoir. The city that receives water from this reservoir wants to determine how much water yield increase can be expected from such a cut. If sufficient water yield increases can be expected, the city may implement a sustainable forest management operation in which portions of their municipal watershed are maintained in clearcut or young-growth conditions.

To provide a conservative estimate of water yield expectations, precipitation and temperature records corresponding to a relatively dry 15-month period were used to perform a water-budget analysis for existing conditions – a mature, mixed hardwood forest (Table 12.2). Soils were clay-loam textured with a plant available soil-moisture content of 164 mm/m. Plot studies indicated that the mature forest had an effective rooting depth of 1.7 m, which means that 279 mm of soil moisture could be used to satisfy ET demands (1.7 m $\times$ 164 mm/m). Based on this water-budget analysis, the 15-month water yield was 242 mm.

To estimate the effects of clearcutting, the same initial conditions were used but the effective rooting depth of the remaining plants was assumed to be 0.8 m, which corresponds to a herbaceous-shrub plant cover, a condition similar to that of a clearcut. The resulting available soil-moisture capacity for the clearcut condition was 131 mm (0.8 m $\times$ 164 mm/m). Water yield for the clearcut condition (Table 12.3) was 390 mm for the same 14-month period. Water yield, therefore, was increased by 148 mm or 273 600 m^3. Of course, one must recognize that the 148 mm increase would be expected at the clearcut site and water can be lost before reaching the reservoir site.

contain modifications of the previously discussed methods. However, in applying any model, it is imperative that the user of the model understands the basic principles and relationships associated with the hydrologic relationships of forest vegetation and forest influences on streamflow response to land-use change. Many of the computer models used by practicing hydrologists were developed for engineering-design purposes or for urban and agricultural landscapes and are not often applicable for forested watersheds. An unwritten rule that should be followed is that a hydrologic model is only as good as the knowledge of the hydrologist applying the model. A more detailed discussion of computer-simulation models and their suitability for wildland watersheds is presented in Chapter 16 .

TABLE 12.2. Water budget for a hardwood-covered watershed before clearcutting

	Year 1									Year 2				
	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May
	(mm)													
Average precipitation[a]	27	4	31	42	36	12	50	120	140	105	90	95	65	20
Initial soil moisture[b]	279	248	163	67	0	0	0	0	100	240	279	279	279	279
Total available moisture	306	252	194	109	36	12	50	120	240	345	369	374	344	299
Potential ET[c]	58	89	127	173	157	107	57	20	0	0	3	13	58	89
Actual ET[d]	58	89	127	109	36	12	50	20	0	0	3	13	58	89
Remaining available moisture	248	163	67	0	0	0	0	100	240	345	366	361	286	210
Final soil moisture[e]	248	163	67	0	0	0	0	100	240	279	279	279	279	210
Water yield[f]	0	0	0	0	0	0	0	0	0	66	87	82	7	0

[a] Average over the watershed for each month of record.
[b] At start of each month. Same as "final soil moisture" of previous month.
[c] Average annual values for the month, as estimated by Thornthwaite's method.
[d] Total available moisture, or potential ET, whichever is smaller.
[e] At end of month. Same as "initial soil moisture" for next month. This value cannot be larger than the available soil-water–holding capacity determined for the watershed, for this watershed 279 mm.
[f] Water is yielded when the remaining available moisture exceeds the water holding capacity for the soil in the watershed (279 mm).

TABLE 12.3. Water budget for a clearcut hardwood forest

	Year 1									Year 2				
	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May
	(mm)													
Average precipitation	27	4	31	42	36	12	50	120	140	105	90	95	65	20
Initial soil moisture	279	248	163	148	148	148	148	148	248	279	279	279	279	279
Total available moisture	306	252	194	190	184	160	198	268	388	384	369	374	344	299
Potential ET	58	89	127	173	157	107	57	20	0	0	3	13	58	89
Actual ET[a]	58	89	46[a]	42	36	12	50	20	0	0	3	13	58	89
Remaining available moisture	248	163	148[a]	148	148	148	148	248	388	384	366	361	286	210
Final soil moisture	248	163	148	148	148	148	148	248	279	279	279	279	279	210
Water yield	0	0	0	0	0	0	0	0	109	105	87	82	7	0

[a] Actual ET is restricted by the available soil-water capacity of the reduced rooting zone (279 mm − 131 mm = 148 mm); the final soil-water content must still exceed 279 mm before any water is yielded.

Forest Management and Streamflow Regimen

Water-resource problems are related to the timing of streamflow discharge as much as to the magnitude of annual water yields. Flow from most watersheds and river basins experience seasonal high and low flows, with floods and drought-induced minimum flows representing two extremes of streamflow that result largely from meteorological events. Solving problems of such streamflow extremes involves a variety of nonstructural- and structural-engineering approaches, as well as people management (refer to Table 1.1 in Chapter 1). Structural solutions include reservoirs and levees. Floodplain management and zoning represent a nonstructural alternative. Vegetation management of upland watersheds and along stream channels should be included as a complement to either approach.

Forests exert a pronounced influence on streamflow yields (as discussed previously) and also on the timing and magnitude of high and low streamflows. By not properly accounting for the effects of forest vegetation on the amount and timing of water yield, watershed management objectives can become compromised (Box 12.3). However, there has long been a myth that forests influence the timing of streamflow by storing water during wet periods and releasing water during dry periods (FAO, 2005). Although forests have no such mechanisms to regulate flows, this myth and similar misunderstandings about the hydrologic function of forests have provided the impetus for forest conservation movements in Europe and the USA (Andréasian, 2004, and others).

Forests are the dominant cover on most of the source areas of headwater streams in the USA. Headwater streams typically comprise 60–80% of catchments and are critical sources of water, sediment, woody debris, organic matter, and nutrients to downstream water bodies

Box 12.3

Water Resource Problems Resulting from Vegetative Changes (Drysdale, 1981)

To develop a wood-based industry on the Fiji Islands, 60,000 ha of *Pinus caribaea* were planted on the country's two largest islands, Viti Levu and Vanua Levu. Plantations were established on the dry, leeward zones of both islands. On Viti Levu, a water supply dam with hydroelectric power stations was developed coincident with afforestation. The project was intended to supply water to the two largest "dry zone" towns on the island for 30 years. Although annual rainfall on the windward sides of the mountains can exceed 4800 mm/year, rainfall during the dry season (May through October) on the leeward slopes varies from 300 to 500 mm with prolonged dry periods. As forest cover replaced mission-grass cover, dry-season streamflow diminished, a cause for concern to the water supply project. Streamflow reductions of 50–60% were observed from watersheds that had 6-year-old pine stands. Greater reductions were expected once pine forests became mature. In this instance, the afforestation project was at cross-purposes with the water resource project.

(MacDonald and Coe, 2007). The following sections examine the extent to which forested watersheds can be managed to help solve problems of stormflow and flooding, low flows, sediment, and water quality.

Effects on Stormflow

Changes in the forest cover on a watershed can impact stormflow peaks, volumes, and timing of stormflows. Such changes are “more observable” when the rainfall or snowmelt-runoff events generating the stormflow are “average” rather than “extreme” because impacts of the change in forest cover are frequently masked when extreme-precipitation events occur. Other than the extreme, stormflow characteristics often change in relation to the severity and extent of the disturbance on a watershed.

Forest Harvesting – Rainfall Regimes. The effects of removing forest overstory on stormflow volumes, peaks, and timing are dependent on several factors, particularly the level of overall disturbance to the forest vegetative cover and soils. The effects of forest harvesting and related disturbances on rainfall-induced peaks and volumes differ from those resulting from snowmelt; therefore, we discuss these effects separately.

While some indicate that rainfall-caused peaks and stormflow volumes increase following forest harvesting (Moore and Wondzell, 2005; Grant et al., 2008), others report variable effect in stormflow characteristics (Table 12.4). The variability in peak-stormflow responses to cutting trees limits our ability to generalize about expected effects. However, much of the variability can be attributed to different combinations of rainfall, the magnitude, antecedent soil-moisture conditions, the proportion of area affected by the tree cutting, and the topography of the watersheds.

An investigation indicating a reduction in peak stormflows following clearcutting the forest on a watershed in British Columbia, Canada (Table 12.4), showed that peak flows were delayed by several hours because of disturbances to the soil that resulted in a “rough” surface with greater retention storage (Cheng et al., 1975). In addition, the velocities of the peak stormflows were reduced by accumulations of debris in the channel following logging.

Annual water yields increased by 10% following clearcutting tropical rainforest in Australia but had little effect on the peak-stormflow parameters (Gilmour et al., 1982). The mechanism of the change in stormflow discharge explains this discrepancy. Soils on the watershed had a hardpan with a reduction in hydraulic conductivity at depths from 0.2 to 0.5 m restricting rapid percolation. The zone of low permeability in these soils – not the vegetative cover – dominated the stormflow process during wet seasons.

Forest Harvesting – Snowmelt Regimes. The increases in water yield caused by forest removal have the potential to also increase flood-producing flows. In addition to increased volumes of snowmelt runoff, the rate of snowmelt can be increased in cleared areas due to increased energy available to melt snow (see Chapter 3) resulting in snowmelt reaching streams quicker from open areas than from forested areas.

Changes in forest cover in cold-continental climates can also affect snowmelt runoff by altering soil-frost conditions and, thereby, runoff efficiency. Soils are generally wetter in the autumn after a clearcutting of trees and they tend to freeze deeper with a greater occurrence of concrete frost. Snowmelt on frozen soil runs off the soil surface rapidly because of the

TABLE 12.4. Changes in rainfall-produced stormflow following forest cover removal

			Changes in stormflow			
Location, climate, zone	Vegetation and soils	Treatment and percent of area affected	Peak (%)	Volume (%)	Timing (%)	References
North Carolina, USA, humid, temperate	Mixed hardwood, moderately deep soils, gravelly loam	Clearcut, 100% (no log removal)	+6	+11	0	Hewlett and Helvey (1970)
New Hampshire, USA, humid, temperate	Mixed hardwood, sandy loam soils, average depth o.s.m.	Clearcut, 100% regrowth prevented by herbicides for 3 years (no log removal)	+100 to +200	+30	0	Hornbeck (1973)
Oregon, USA, humid, coastal	Conifers, shallow to deep sandy soils	Commercial clearcut, 100% (logging, road construction)	+100 or less	+10 or less	0	Harr et al. (1975)
Minnesota, USA, cold, continental	Mixed hardwood, glacial till soils, medium depth	Commercial clearcut, 70%	+170	+200	0	Verry et al. (1983)
British Columbia, Canada, humid, temperate	Conifers, gravelly sandy loam soils from glacial till	Commercial clearcut, 100% (logging)	−22	NR[a]	Delayed	Cheng et al. (1975)
North Queensland, Australia, humid, tropics	Rainforest, deep clay soils, shallow hardpan	Clearcut, 100% (pasture development)	0	0	0	Gilmour et al. (1982)

[a] NR means not reported.

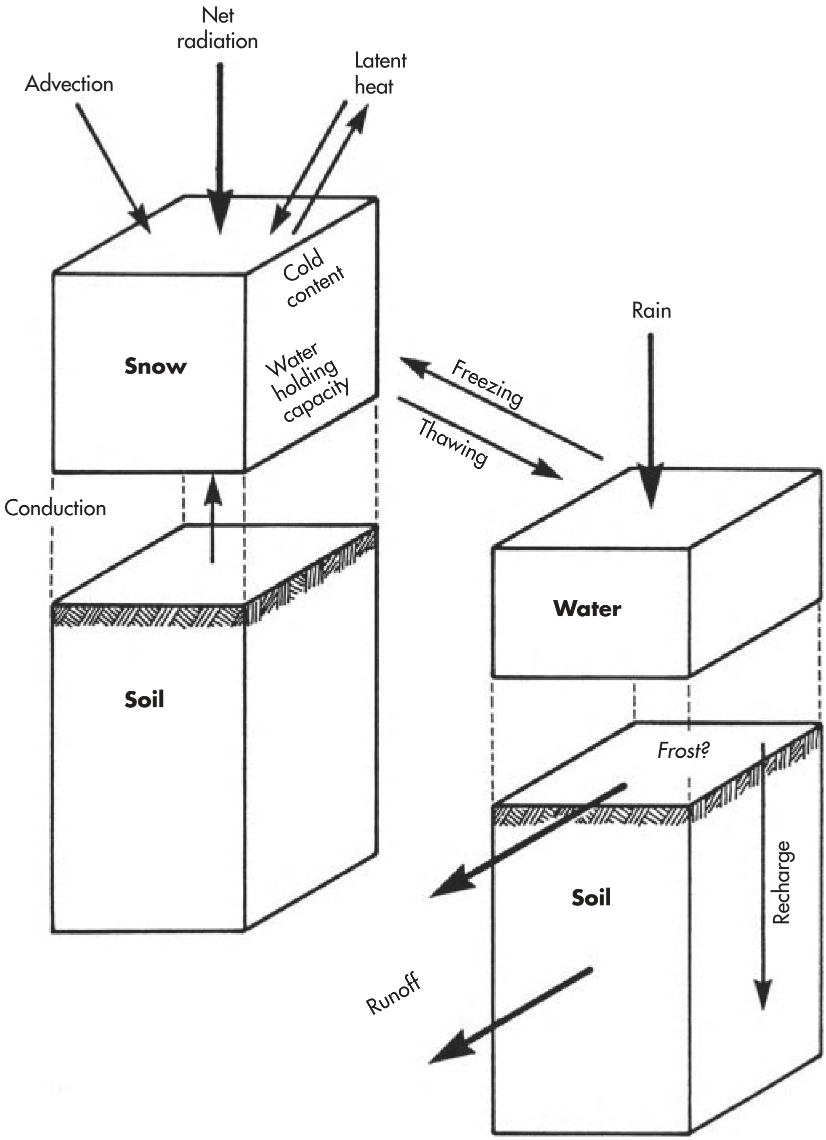

FIGURE 12.6. Factors that affect snowmelt and runoff from a snowpack

reduced infiltration capacity of the frozen soil (see Chapter 5). The factors that collectively lead to greater snowmelt-runoff efficiency are illustrated in Figure 12.6.

The timing of snowmelt runoff can be affected by changes in forest cover. In locations with minimal topographic relief, forest cover can be manipulated to alter the timing of snowmelt from different parts of a watershed or river basin. This effect has been observed in Minnesota where a mature mixed-aspen stand on a watershed was cleared only partially in 2 successive years (Verry et al., 1983). When forest cover from one-half of the watershed was cleared, the snowmelt-runoff peak was reduced (the year 1971 in Fig. 12.7). When more

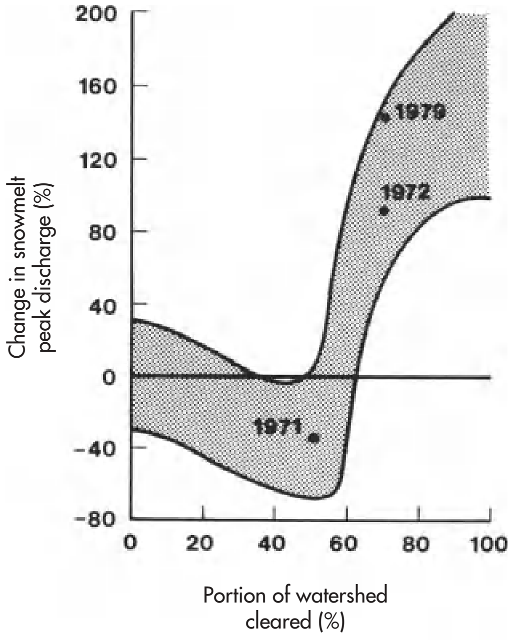

FIGURE 12.7. Relation between the portion of a watershed clearcut and the change in annual snowmelt peak discharge compared to a mature stand of hardwoods in Minnesota (from Verry et al., 1983)

than 70% of forest cover was cleared the next year, snowmelt peak discharge was nearly doubled (1972 in Fig. 12.7). The desynchronization of melt between open and forested areas accounts for the desynchronization of streamflow peaks. These effects can persist for several years (through 1979 in Fig. 12.7). Other areas with similar climate, vegetation, and topography might experience a similar response to forest clearing.

Clearing patches of forest cover in mountainous areas would not be as effective in reducing peak stormflow as clearing forest patches in flat terrain (Kattelmann and Ice, 2004). Snowmelt runoff from mountainous watersheds is naturally desynchronized because of differences in elevation, slope, and aspect.

In general, peak discharges from snowmelt runoff have increased after forest removal. However, exceptions have been observed in the coastal range of the Pacific Northwest. Harr and McCorison (1979) reported a 32% reduction in snowmelt-peak size following clearcutting in Oregon and suggested that condensation–convection melt from snow on tree crowns occurred at a higher rate than snowmelt at ground level where most of the snow would be after clearcutting. Such conditions would be encountered in areas with transient snowpacks along coastal areas where snowmelt rates can be accelerated by advective energy from the ocean.

Peak stormflows following snowmelt runoff on a 289 ha watershed in the subalpine forests on the Frasier Experiment Forest increased 20% after clearcutting the forest in a series of strips that were parallel to the contours (Stednick and Troendle, 2004). The increase in peak stormflows was correlated positively with the snowpack-water equivalents

TABLE 12.5. Effect of wildfire on annual peak flows

Location	Vegetation type	Increase %	References
Eastern Oregon	Ponderosa pine	45	Anderson et al. (1976)
Central Arizona	Ponderosa pine	500–1500	Rich (1962)
Cape Region of South Africa	Monterey pine	290	Scott (1993)
Southwestern USA	Chaparral	200–45,000	Sinclair and Hamilton (1955) and Glendening et al. (1961)

on the watershed. The time to peak also occurred 1 week earlier in comparison to that on an adjacent uncut watershed.

If forest cover is removed over large areas within a watershed the combination of more rapid snowmelt and greater snowmelt-runoff efficiency can lead to more frequent downstream flooding. Rain-on-snow events, particularly when soils are frozen, result in some of the most severe floods in North America. However, the role of forest cover in mitigating floods is limited.

Effects of Wildfire. When forest cover and accumulations of organic matter on the soil surface are destroyed by a wildfire, reductions in interception and ET losses generally result with decreases in infiltration rates, increases in overland flows and often an increase in peak stormflow. The largest increases are associated with intense- and short-duration rainfall inputs, slope steepness of the burned watershed, and the possible formation of water-repellent soils after the burning (DeBano et al., 1998; Moody and Martin, 2001). However, the magnitude of these increases is variable (Table 12.5). Postfire stormflow events with excessively high peaks are often characteristic of flooding regimes.

The largest postfire-peak stormflows in the western USA have been measured on small watersheds at the headwaters of river basins (Biggio and Cannon, 2001). For example, the peak stormflow from a 25-ha watershed at the headwaters of the Little Colorado River burned by a high-severity wildfire shortly after cessation of the Rodeo-Chediski Wildfire in northern Arizona – one of the largest known wildfires in the state – was 6.57 m^3/s or 2235 times larger than that measured before the wildfire (Fig. 12.8). This peak stormflow represented the largest postfire-peak stormflow measured in the forests of the southwestern USA (Ffolliott and Neary, 2003). Fortunately, this and the other peak stormflows from the larger area burned by the wildfire did not result in excessive flooding downstream.

Stormflow regimes from burned watersheds often respond to rainfall inputs faster than watersheds supporting a protective forest cover, producing streamflow events where time-to-peak is earlier (Neary et al., 2005). Timing of snowmelt runoff in the spring can also be advanced by the occurrence of a wildfire (Helvey, 1973). The earlier snowmelt is initiated by lower snow albedo caused by blackened trees and increased surface exposure with elimination of a forest cover.

High-severity wildfires that reduce the time-to-peak also increase the possibility of flooding, at least for first-order streams. For example, the historic Tillamonk Burns of 1933 in Oregon increased peak stormflows from two burned watersheds by almost 1.5

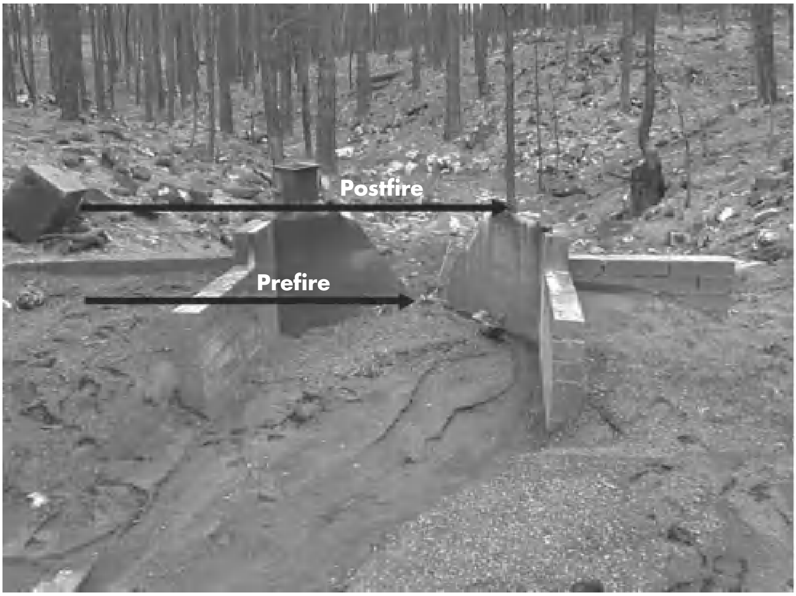

FIGURE 12.8. Streamflow discharge peak-flow stage shortly after cessation of the Rodeo-Chediski Wildfire in a ponderosa pine watershed in Arizona; prefire peak-discharge stage noted (For a color version of this photo, see the color plate section)

times because of the loss of protective forest covers and extensive exposures of bare soil (Anderson et al., 1976), increasing the potentials for downstream flooding to occur. First-order watersheds burned by wildfire respond to rainfall events faster than snowmelt runoff and, in doing so, can produce more "flashy" flood flows. As a consequence, warning times are likely to be reduced with the flood flows causing damage to property and, at times, a loss of human life. Large-scale postfire flooding is not often the case, however, as witnessed by the flows following the Rodeo-Chediski Wildfire mentioned earlier.

Effects on Baseflow and Low Streamflows

Dry-season streamflow is a concern to many people because it often coincides with periods of greatest need. Usually dry-season flows are sustained by groundwater flow when precipitation is lacking and when groundwater and reservoir storage are needed to provide water for irrigation and municipal requirements. Pollutants become concentrated during periods of low streamflow and streams become more sensitive to perturbations and temperature fluctuations (Hornbeck et al., 1997). Low flows can place aquatic ecosystems under considerable stress. Because of the consequences of low flows, management of resources should be aimed at maintaining streamflows during dry periods or at least not diminishing low flows further.

The relationship between forests and dry-season streamflows needs to be clearly understood so that decisions on forest management do not diminish dry-season flows as the example of Fiji in Box 12.3. By increasing forest cover to enhance dry-season flow in this case was a mistake. Theoretically, forests in areas with deep soils and high rainfall can be important recharge areas for groundwater, assuming that the amount of rainfall infiltrating and percolating below plant roots will exceed the amount of water that is consumed by ET. In the Fiji example, this was not the case. The results of many experimental watershed studies indicate that because of their high ET, forests tend to diminish rather than enhance flows during the dry season.

Studies that have focused on the effects of forest removal on low streamflows have shown an initial increase in low flows shortly after the tree-cutting activities (Keppeler and Ziemer, 1990; Hornbeck et al., 1997; Swank et al., 2001). However, the observed increases in low flows are often less than 10 years because of the recovery of vegetation resulting in increases in interception capacity and ET rates. If forest cover were replaced with shallow-rooted herbaceous vegetation with lower interception and lower ET rates, one might expect to see a more sustained response of increased dry-season flows.

Wetter soil conditions resulting from reduced ET losses can cause a watershed to be more responsive to small rainfall amounts and, in doing so, lead to a longer period of soil-water drainage and subsurface flow. However, because groundwater is most important in sustaining dry-season flows, continued precipitation inputs are needed to sustain the flow. Changes in vegetative cover would not be expected to substantially alter low flows resulting from extended dry periods or droughts in any case.

Streamflow resulting from snowmelt runoff is usually characterized by long recession flows. Since forest cover can be manipulated to affect snowpack accumulation and melt, it has been suggested by some watershed managers that snowpacks can be “managed” to enhance low streamflows in dry seasons. Kattelmann et al. (1983) stated that the greatest contribution watershed management can make in meeting future water demands in California might be achieved by manipulating forest cover to delay and extend snowmelt runoff. A significant delay in streamflow response could be beneficial for the operation of reservoirs. Unfortunately, delayed and increased streamflows caused by vegetative manipulations normally extend over short periods and would not affect flows during lengthy droughts.

Baseflow is likely to decrease when the condition of a watershed deteriorates as a result of a wildfire or other disturbance with more of the excess rainfall or snowmelt runoff leaving the watershed as overland flow. Perennial streams become ephemeral in extreme situations. However, the occurrence of a wildfire in combination with other land-use activities on a watershed can increase baseflow in other situations. While the mechanisms involved were unclear, Berndt (1971) observed immediate increases in baseflow following a forest fire on a 571-ha watershed in eastern Washington. This increase in baseflow persisted above prefire levels for 3 years after the fire occurred.

Water Quality

Considering all water-quality indicators, the streamflow from undisturbed-forested watersheds is normally of the highest quality. This characteristic of forests was one of the justifications used in developing the national forest system in the USA – to sustain the production of high-quality water. Therefore, forest management practices and land use on forested watersheds undergo close scrutiny to sustain high-water quality and to mitigate any detrimental impacts on the quality of streamflow. Even though flows from forested watersheds must comply with the established water-quality standards for the region, as with any water bodies, one gets the sense that water flows from forests are held in the highest standards. The following summarizes the physical, chemical, and biological characteristics of water flow from forests and management options to protect and sustain high levels of water quality.

Physical Characteristics

Sediment, turbidity, water temperature (thermal pollution), and dissolved oxygen are key physical characteristics of water quality of particular interest in forested streams.

Sediment. As discussed in preceding chapters, forested-headwater watersheds are important sources of the flow of water and sediment. Sediment in streamflow occurs naturally and is not a pollutant unless levels exceed "background levels" that would be expected to occur naturally – without human interventions. Background levels of sediment are not constant and do not conform to water-quality standards that establish threshold levels that are not to be exceeded. Therefore, monitoring of sediment is an essential ingredient in managing watersheds in an attempt to maintain and not exceed naturally occurring levels of sediment.

Annual sediment yields from uncut and clearcut watersheds on the Hubbard Brook Experimental Forest in New Hampshire averaged 40 kg/ha for several decades (Hornbeck and Kochenderfer, 2004). These sediment yields were among the lowest measured in the USA. However, they were "highly variable" from year-to-year depending on the frequency of "unusually" large rainstorm events in a year. While sediment production had increased 10- to 30-fold in the first few years after completion of cutting and skidding operations, it then steadily declined with the regrowth of vegetation.

The process of timber harvesting – felling, limbing, and bucking of trees – generally do not contribute to significant increases in sediment loading in streams. However, the associated activities such as the skidding operation that moves cut trees to a roadside landing can significantly increase soil erosion that in turn increases sediment in the stream channel (Stednick, 2000). Ground skidding causes the most erosion, and aerial skidding, the least.

Sediment yields from watersheds in the southern Appalachian Mountains located in the southeastern USA were unchanged after tree-cutting activities when proper road construction is followed (Swank and Tilley, 2000). Similarly, the changes in the sediment production from watersheds in the central Rocky Mountains were generally small and short-lived following timber-harvesting activities when the best management practices are followed (Stednick, 1987). The largest portion of the sediment loads from these watersheds were derived from streambank erosion and channel degradation while suspended-sediment concentrations are normally low.

Sediment yields following a wildfire vary considerably depending on the intensity of the fire, the weather conditions encountered, the geomorphic features of the burned watershed such as geology, soils, and topography, and the recovery postfire vegetation. For example, Noble and Lundgreen (1971) reported that the annual sediment production averaged of 5.7 mg/ha from a burned watershed of 365 ha after a wildfire on the South Fork of the Salmon River in Idaho. This magnitude of sediment was approximately 7 times greater than that from nearby unburned watersheds. Elsewhere, annual sediment production increased from 100 to 3800 kg/ha following a wildfire in coniferous forests on the eastern Cascade Mountains of Washington (Helvey et al., 1985). DeBano et al. (1996) reported that the sediment yields following a low-severity wildfire on a watershed in the ponderosa pine forests of northern Arizona returned to prefire levels in 3 years but that it took 7 and 14 years, respectively, for postfire sediment yields from moderately and severely burned watersheds to recover.

Livestock grazing within the carrying capacity of a forested watershed does not normally change the delivery of sediment to a stream channel with the following exceptions. Intensive grazing on steep terrains of fragile soils can increase soil erosion and, as a result, increased sediment yields. Increases in sedimentation also occur when the livestock graze in riparian corridors causing streambank erosion and sediment depositions directly into stream channels (Magner et al., 2008). Therefore, livestock grazing should be excluded on steep slopes and managed to retain healthy riparian vegetation and protect streambank root depth and density.

Compacted road surfaces, cutslopes above and fillslopes below roads, and roadside ditches are often exposed and, as a result, can increase soil erosion and surface runoff that transports sediment into streams. Overland flows of water on road surfaces and within roadside ditches can also concentrate the soil moisture on steepened cutslopes and fillslopes to increase the occurrence of landslide erosion (Wemple et al., 2001). Even when these corridors are abandoned, they can continue to be a source of sediment in streamflow runoff.

Turbidity. Turbidity is one of the most visible of the water-quality responses to timber harvesting or a wildfire and often is related to high levels of suspended sediment and organic material in the stream. As indicated in Chapter 11, turbidity can be used as a surrogate for concentrations of suspended sediment if there is an established correlation between the two parameters.

Impacts of timber-harvesting operations and silvicultural cuttings on turbidity are minimal when these interventions are carefully planned and implemented. For example, there was an insignificant impact on turbidity following the clearcutting of a watershed on the Hubbard Brook Experimental Forest (Hornbeck and Kochenderfer, 2004). However, flows of turbid water can increase as the consequence of a wildfire.

Increased turbidity following a wildfire is generally the result of the suspensions of ash and silt- to clay-sized particles in the streamflow (Neary et al., 2005). Turbidity levels are also affected by the steepness of the burned watershed (Landsberg and Tiedemann, 2000). For example, the turbidity of overland flows of water from burned watersheds in the southwestern USA is higher on slopes in excess of 20% than on slopes of lesser steepness with much of this turbid water ending up in postfire streamflows. However, other activities on these watersheds including the burning of the slash from earlier timber harvesting or seeding burned sites with herbaceous plants should also be considered when evaluating these effects.

Water Temperature. The temperature of streamflow is a function of the temperature of inflowing water, the inputs and outputs of radiant and conductive energy, and the heat that is lost from the stream as a result of evaporation. Riparian vegetation influences all of these exchanges. Beschta et al. (1987) described the relationships between either the cutting of trees or the occurrence of wildfire that eliminates or greatly reduces the vegetation in riparian corridors of the Pacific Northwest and the resulting increases in stream temperatures. Summer temperatures of streamflow from watersheds in the southern Appalachian Mountains increased from 20°C to about 25°C following the clearing of the forest cover

(Jackson et al., 2004). However, the water temperatures later returned to pretreatment levels with the increased shading provided by the regrowth of riparian vegetation.

Increases in the temperature of streamflow water because of the elimination of riparian vegetation by a wildfire can also result in thermal pollution that in turn increases the biological activity in a stream (DeBano et al., 1998). However, streams with significant deep groundwater inflow will buffer the loss of vegetation better than streams which lack deep groundwater resurgence. When thermal pollution occurs, increases in biological activity can place a greater demand on the dissolved-oxygen content of the water, an important water-quality characteristic from a biological perspective.

A stream transports heat from one site to another by changes in temperature. Therefore, the effects of eliminating or reducing riparian vegetation on water temperature can impact on downstream water uses that are dependent on temperature (Quigley, 1981). Among attributes that help to determine the contributions of riparian vegetation in mitigating thermal pollution are stream width and orientation, the distance from riparian vegetation to the stream, and the canopy characteristics of the riparian vegetation.

Dissolved-Oxygen Levels. Changes in land-use activities that reduce the dissolved-oxygen levels of a stream can also decrease the habitability of the stream for fish and other aquatic life. For example, an increase in nutrient loading depletes dissolved-oxygen levels by stimulating algal blooms. Detrital organic materials consume oxygen as they decay. The cumulative effects of all factors on the temperature and oxygen regime of streams should be recognized by watershed managers.

Changes in sediment yields, turbidity levels, water temperature, and dissolved-oxygen levels are all generally less following a prescribed burn than a high-severity wildfire because of the lower fire intensities.

Chemical Constituents

The removal of trees through timber harvesting or a wildfire, changes in sediment transport of nutrients and heavy metals, and the use of pesticides, fertilizers, and fire retardants can affect the chemical characteristics of streamflows.

Either the cutting of trees or a wildfire can affect the ionic balances of streamflow. The dissolved-chemical concentrations in a stream after cutting trees or a fire are largely a function of the biotic and abiotic characteristics of the ecosystem of the watershed. For example, leaching rates are influenced by the form, amount, and intensity of precipitation events. The composition and densities of plant species also influence the rates of nutrient uptake (Stednick, 2000). Regrowth rate of vegetation after a disturbance can control the recycling of nutrients following the disturbance to a large extent.

Nutrients. There is often an increased loss of nutrients following timber harvesting on watersheds with well-drained soils. For example, clearcutting hardwood forests on the Hubbard Brook Experimental Forest in New Hampshire resulted in increases of 57 kg/ha in inorganic nitrogen (N), 71 kg/ha in calcium (Ca), and 15 kg/ha in phosphorus (P) in the year immediately after cuttings (Martin et al., 1986). The largest increases were observed in the second year after the cutting with concentrations of most nutrients returning to preclearcutting conditions by the end of the fourth year. Even if concentrations of nutrients in streamflow do not increase, increases in overland flows of water can increase the total

Box 12.4

Effects of Clearcutting Trees and Site Preparation on Nutrient Export from Small Watersheds in Southeastern USA

Clearcutting of loblolly pine followed by site preparation with roller choppers and planting caused nutrient export to increase, but the increase was small and did not last beyond 2 years (Hewlett et al., 1984). The concentrations of nutrients in streamflow did not increase following clearcutting, but water yield increases of 10–20 mm over 2 years flushed more nutrients from the watershed than a paired (uncut) control. Nitrate-N loading increased 0.3 kg/ha for 2 years following harvesting. The maximum nutrient flushing following timber harvesting was less than 0.5 kg/ha per month for P, K, Ca, Mg, and Na; these were short-lived and diminished as regrowth occurred. Such nutrient losses result in neither soil fertility losses nor stream eutrophication.

Experimental watersheds at the Coweeta Hydrologic Laboratory in North Carolina experienced larger nutrient losses following hardwood clearcutting than those reported in above. Net losses of 6.2 kg/ha and 3.6 kg/ha of N were measured for the first and second years following clearcutting (Swank and Waide, 1979). Again, the losses reported were not considered to be a threat to the quality of streamflow.

nutrient loading of water bodies. However, this loading is rarely significant when timber harvesting and silvicultural cuttings are planned to minimize site disturbances and lead to a rapid establishment of replacement forest stands.

Similar levels of nutrient export have been observed following tree-cutting and site-preparation activities on watersheds in the southeastern USA (Box 12.4). The reported losses of nutrients did not generally adversely impact the quality of streamflow.

Concentrations of nitrate nitrogen (NO_3) in a stream, one of the most mobile nutrients in a disturbed forest ecosystem, often increase after a timber harvest (Binkley and Brown, 1993; Martin et al., 2000). However, the elevated NO_3 concentrations generally return to pretreatment levels with regrowth of forest vegetation (Likens and Bormann, 1995; Swank et al., 2001). Concentrations of P can also increase following tree-cutting activities or when postfire-stormflow events transport P to the stream system (Meyer and Likens, 1979). Riparian buffer strips can reduce particulate P attached to sediment but are less effective in removing dissolved phosphorous.

Burning slash left after timber harvesting or silvicultural cuttings can produce an even greater and often more rapid release of ions from litter and mineral soil than the tree-cutting activity itself. This increased release of ions is largely due to a breakdown of organic materials into a soluble form and making the nutrients easily removable by leaching (Landsberg and Tiedemann, 2000). While this process can lead to an increase in the total loss of nutrients, the increase is only temporary in many instances. Volatilization of nitrogen and sulfur by fire also results in losses from the watershed.

Moderate-to-severe wildfires in coniferous forests of the western USA have had minimal effects on the concentrations of dissolved Ca, magnesium (Mg), sodium (Na), and bicarbonate (HCO_3) in streamflow (Helvey et al., 1985). The occurrence of “light” rainfall events following a wildfire can dissolve and leach the ash constituents into permeable soil where they can be adsorbed in the soil–ion exchange complex instead of being washed into the stream. However, if there is a significant overland flow of water from high-intensity rainstorms following severe wildfires, large quantities of soluble ash can move into streams, a process that can occur for the first few years after the wildfires. But, as vegetation regenerates, the loading of dissolved chemical in the stream is likely to return to prefire levels.

Transport of Nutrients and Heavy Metals in Sediment. Losses of nutrients and heavy metals from watersheds are usually measured as concentrations of dissolved ions in streamflow. However, a source of nutrient and heavy metal loss that is often ignored is that adsorbed by sediment. Sediment has been found to transport relatively high levels of nutrients and heavy metals in large rivers (Angino et al., 1974; Potter et al., 1975) but less is known about nutrient and heavy metal transport by sediment from upland watersheds. One exception is the study described below.

Analyses of sediment from watersheds in the southwestern USA with limestone, sandstone, granite, and basaltic geologies and supporting conifer forests have shown that limestone is high in Ca and potassium (K) while basalt is high in Na (see Chapter 11). Mg is highest in the sand fraction (0.061–200 mm) of basalt and the clay-silt fraction (<0.01 mm) of limestone. Sandstone often has the lowest concentrations of these elements and granite is frequently intermediate. The nutrient adsorbed to the sediment particles are indicative of the geologic formation of a watershed while the type of forest cover affects the organic-matter content, total P, and levels of extractable nutrients of the sediment (Gosz et al., 1980).

Variations in the levels of zinc (Zn), iron (Fe), copper (Cu), manganese (Mn), lead (Pb), and cadmium (Cd) in the streamflow from the watersheds studied by Gosz and colleagues were correlated with sediment concentrations. Furthermore, sediment yields from the different geologies of the watersheds had increasing concentrations of heavy metals in the order of sandstone, granite, limestone, and basalt. Therefore, land-use practices that increase sediment yields also have the potential to increase nutrient and heavy metal losses from a watershed.

Pesticides and Fertilizers. Applications of pesticides to eliminate competing vegetation are often used in site preparation before planting trees. In contrast to the mechanical methods of site preparation, the use of pesticides minimizes off-site soil loss, eliminates on-site soil and organic-matter displacement, and prevents the deterioration of soil physical properties (Michael and Neary, 1995). Residue concentrations of such pesticides as 2,4-D, hexazinone, picloram, sulfometuron methyl, metsulfuron methyl, triclopyr, and imazapyr tend to be low, generally less than 0.1 mg/L (0.1 ppm). Furthermore, these concentrations do not persist for long periods of time and become problems when applications are made directly to ephemeral-stream channels or streams themselves. Concentrations greater than 2 mg/L (2 ppm) would be necessary to detrimentally affect stream flora. Pesticide applications that follow regulatory guidelines should not impair water quality.

The migration of pesticides into streams has been documented while their movement into groundwater aquifers is less well known. Theoretically, the movement of pesticides from a stream into a groundwater aquifer should reduce their initial concentrations for several reasons (Michael, 2000). Infiltrating pesticides must pass through several physical barriers (layers) before reaching the groundwater. Pesticides become degraded, diluted, and metabolized as they pass through each barrier. Surface water is a medium for dilution, hydrolysis, and photolysis. Aquatic vegetation can metabolize pesticides. Microbes associated with particulate-organic matter found naturally in streams also metabolize pesticides.

Application of nitrogen-based fertilizers in forest stands can increase the nitrogen concentrations in streamflow. Urea and ammonia levels often remain below levels of concern in most situations but NO_3 levels can peak at high concentrations. However, studies in the Pacific Northwest have shown that the risk of NO_3 pollution is small when fertilizing Douglas-fir stands (Bisson et al., 1992).

Fire Retardants. Fire retardants applied to suppress wildfires can also result in adverse-environmental impacts. For example, ammonium-based and nitrogen-containing retardants have the potential to degrade water quality and detrimentally impact aquatic organisms (Norris and Webb, 1989). Of concern is the possibility of high concentrations of these and the other toxic chemicals in fire retardants in streams after their application (Neary et al., 2005). Careful planning and operational control of applying fire retardants in suppressing wildfires are necessary for minimizing their effects on streams and their biota.

Biological Characteristics

Streams in undisturbed-forested watersheds generally exhibit high levels of biological diversity and low levels of pathogenic bacteria and protozoa as discussed in Chapter 11. Biological diversity and pathogenic bacteria and protozoa are particularly susceptible to major disturbances in the riparian areas of forested watersheds – areas that have received the greatest attention by watershed managers.

Pathogenic bacteria and protozoa in surface water are a concern because of their potential to cause diseases. Tree-cutting activities and forest fires do not generally affect the occurrence of either pathogenic bacteria or protozoa in surface water but grazing by livestock can cause increased fecal-coliform bacteria in streams, especially when the livestock graze in riparian corridors. Therefore, grazing of livestock and other ungulates in these streamside corridors should be controlled to minimize impacts on the bacteriological quality of water.

Concentrations of bacterial groups in streamflow can also be related to the characteristics of the stream. For example, bacterial counts in the streams of the Rocky Mountains appear to be dependent largely on the repeated flushing effect of surface runoff from rainstorms and snowmelt-runoff events. Low counts prevail when the water temperature approaches freezing, while high counts appear in rising and peak streamflows resulting from late spring rainfall and snowmelt-runoff events. High bacterial counts are common in the summer period of warm temperatures but decline with cooler temperatures in autumn. Maintaining buffer strips containing plant litter, duff, and other organic material generally reduces surface-runoff loading of bacterial groups from entering streams.

Roads and Water Quality

Surface runoff and seepage from roads can contain elevated levels of sediments and chemical pollutants. These corridors often continue to be a source of pollution even after they have been abandoned.

Sediment and Roads. The impacts of roads on soil erosion and consequent sedimentation can be greater than all of the other land-use activities on forested watersheds combined (Megahan and King, 2004). Assuming that the rates of soil erosion can be proxies for amounts of sediment production, the sediment yields originating on the bare slopes of road surfaces are often "high" immediately after roads have been constructed but then decline as the exposed surfaces stabilize and revegetate. Soil erosion reductions of 90% or more are common as a road ages (Ketcheson and Megahan, 1996). Nevertheless, road surfaces remain a main source of sediment as long as vehicular traffic or maintenance prevents the establishment of protective vegetation (see Chapter 8).

Burroughs and King (1989) identified the cutslope, the roadway, and fillslope as three significant sources of sediment in the Intermountain Region of the USA and suggest applications of mulch, seed, and sod as mitigation measures for each of these sources. Elsewhere, Luce and Black (1999) concluded that the roadway itself and roadside ditch were the main sources of sediment in the Coastal Range of Oregon. Slopes and channels downhill from a road can also be the site of deposition of soil eroded by the road or (at least) a major source of sediment from a segment of the road.

More sediment originates from road-caused landslides and timber harvesting than surface-soil erosion in areas of persistent high rainfall on steep terrain such as the Coastal Mountain Range of Washington and Oregon. Beschta (1978) reported that road-construction activities and timber harvesting in high rainfall and steep areas of the Coastal Range of Oregon caused sediment yields to increase from about 100 mg/km^2 before timber harvesting to 140 mg/km^2 with most of this increase attributed to soil-mass movement. In a more recent study in Oregon, Elliott (2000) suggested that sediment loads originating from landslides in previously undisturbed areas of a watershed were similar to those in areas of a watershed with a road system. While surface erosion is a source of sediment almost every year, landslides tend to contribute large amount of sediment in only wet years and little or no sediment in dry years. This phenomenon also appears to be true for cohesive river bluffs in the Midwestern USA (Day et al., 2008).

Mitigation techniques to reduce the soil erosion originating on road systems (Burroughs and King, 1989; Megahan and King, 2004) include

- surfacing the corridors with gravel;
- decreasing the spacing of cross drainages;
- locating roads away from a stream and limiting their gradient;
- applying straw to and planting tree seedlings on cut and fill slopes on newly constructed road systems; and
- seasonal closure of roads in wet seasons.

Soil erosion on abandoned-road surfaces often drop to background level as the density of vegetation increases (Rummer et al., 1997). Much of the eroded soil from almost any corridor can be deposited below the corridor with proper erosion control and, therefore, never reaches a stream.

Preemptive approaches to mitigating the impacts of roads, trails, and other corridors on sediment and related constituents are better options than attempting to control pollutants as described above. Application of one or more of the following general guidelines offered by Megahan and King (2004) can reduce potential erosional and sedimentation impacts:

TABLE 12.6. Pollutants observed in runoff from road surfaces

Pollutants	Comments	References
Cadmium, copper, lead, zinc	Treated in wetlands	Mungur et al. (1994)
Highway deicing salt	Sodium adsorbed in soil	Shanley (1994)
Polycyclic aromatic hydrocarbons	Altered aquatic communities	Boxall and Maltby (1997)
Total petroleum hydrocarbons, lead, zinc	Reduced by vegetation	Ellis et al. (1994)
Heavy metals, petroleum hydrocarbons, pesticides, sediment, nutrients	Treatment ponds remove up to 95% of pollutants	Karouna-Renier and Sparling (1997)

Source: Dissmeyer (2000).

- recognize and avoid high-erosion hazard areas on a watershed in planning for the layout of the corridors;
- minimize the amount of the watershed landscape disturbed by roads, trails, and other corridors; and
- minimize the number of stream crossings and maximize the slope-travel distance to the streams.

Chemical Inputs from Roads. A number of chemical pollutants have been identified in surface runoff from road surfaces (Table 12.6). Some of these pollutants originate from the road material itself, some chemicals occur in the soil and rock on the site and are released during construction, and some originate from vehicles traveling on the roads (Elliot, 2000). Vehicular traffic and road surfacing can contribute undesirable cations, hydrocarbons, and metals to water. Cations released from a road can have a buffering effect on runoff acidity.

Chemical pollutants originating on road surfaces can be trapped in natural or constructed wetlands or by maintaining forest strips as buffers between the road and stream system. Another control method is the construction of a partial-exfiltration trench that filters out the suspended solids carrying undesirable hydrocarbons and other pollutants from road surfaces (Sansalone and Buchberger, 1995). Ultimately, multiple approaches can be needed to decrease chemical loading from road surfaces. For example, a detention basin might "catch" the first flush of pollutants in the runoff while a filtration system reduces the chemicals in surface runoff from larger storms. However, detention basins and similar structures built to contain surface runoff from roads can themselves become sources of pollution through seepage into the soil or other forms of hydraulic or structural failure (Grasso et al., 1997). Repeated cleaning of the structures is required in this case.

WOODLANDS

The terms *forests* and *woodlands* are often used interchangeably. However, woodlands and savannas usually occur as transitional areas between forests and grasslands or deserts and have the following characteristics:

- They have more open tree canopies usually with grass understory with smaller and fewer trees per unit area than in the denser forests of larger trees.

Figure 12.9. Pinyon-juniper woodlands represent the most extensive woodland type in the western USA

- Because they occur in drier climatic zones than forests, fewer perennial streams originate in woodlands than in the forests and those that do are often intermittent or ephemeral streams of low flows.
- A characteristic of most of the woodlands is a frequent occurrence of a soil-moisture deficiency that must be "satisfied" before streamflow is initiated.
- Occurring at lower elevations with less annual precipitation than forests, woodlands are commonly used for grazing livestock and managed for wildlife with little potentials for increasing water yields through modifications of tree overstories.

Globally the largest woodlands and savannahs that support wildlife and domestic grazing animals occur in Africa and in the Cerrado woodlands and savannahs in eastern South America (see website http://wwf.panda.org/about_our_earth/ecoregions/ cerrado_woodlands_savannas.cfm).

From a watershed management standpoint, these ecosystems are not managed for water production but are important components of many watersheds in dryland regions of the world and require special attention in terms of sustaining habitat and forage production under grazing pressures. Excessive grazing can lead to soil erosion and sedimentation in these ecosystems. Most common in the western USA are pinyon-juniper (Fig. 12.9) and oak woodlands that have been principally managed as winter grazing lands when snowpacks cover grasses in higher elevations.

Water Yield

In the western USA there is little opportunity to reduce ET losses where the annual precipitation is less than 450 mm and its total is exceeded by the PET (Hibbert, 1983). Hibbert reasoned that precipitation amounts below this value are likely not to penetrate far enough into a soil body to influence the storage of moisture that in turn can impact on water yields. Annual precipitation less than Hibbert's threshold are found throughout most woodland ecosystems of the West.

Because of their importance for forage production, pinyon-juniper woodlands have undergone mechanical and chemical methods of removing trees to increase the production of understory herbaceous plants by eliminating soil-moisture competition with tree overstories. Early watershed managers thought that these same methods of conversion might also be used to increase streamflow discharges by replacing the high water-consuming trees with less water-consuming herbaceous plants (Clary et al., 1974). Evaluations of this possibility were subsequently made in the southwestern USA and California. However, conversion of pinyon-juniper woodlands to herbaceous plants by mechanical methods has had little effect on streamflow volumes (Roundy and Vernon, 1999; Kuhn et al., 2007; Ffolliott and Stropki, 2008).

Killing trees with herbicides has also been investigated as a means of converting pinyon-juniper overstories to understories of herbaceous plants in northern Arizona. The purpose of the herbicidal treatment tested was to lower ET losses by killing the trees but leaving them standing to reduce the frequently occurring desiccating winds and intensive-solar radiation to control evaporation losses. Application of herbicides is the only known conversion treatment to significantly increase water yields in the pinyon-juniper woodlands (Baker, 1984; Baker and Ffolliott, 1999). While statistically significant, this increase was less than 13 mm in absolute terms with little promise for enhancing water flows. Furthermore, in general the use of herbicides for watershed management purposes is currently limited by environmental regulations.

Kuhn et al. (2007) investigated the impacts of large-scale removal of juniper trees on streamflow volumes for the larger tributary watersheds in the Klamath River Basin of northern California and concluded that there was little potential for increasing water yields even with the complete removal of juniper trees. Similar efforts in oak woodlands on the lower western slopes of the Sierra Nevada Mountains in eastern California showed no change in water yields after clearcutting oak overstory (Epifanio et al., 1991).

Woodlands and Streamflow Regimen

In contrast to upland forests, runoff in intermittent or ephemeral streams in the woodlands tends to be flashier because of the dominance of surface runoff. High-intensity and localized summer thunderstorms common in these woodlands can produce high-stormflow peaks, particularly on watersheds supporting tree overstories that are so dense that the underlying soil surface is bare of herbaceous plants. Rainfall runs off these bare soils quickly as overland flow. Because of the inherent surface-runoff conditions in these woodlands, conversions to herbaceous cover have had limited impacts on stormflow peaks and volumes and on low flows of the few streams that originate in the woodlands.

Water Quality

The processes that impact on the quality of streamflows in the pinyon-juniper and oak woodlands are generally similar to the processes impacting on the quality of water flowing from forest ecosystems. Of greatest concern is management to prevent excessive soil erosion and sediment delivery downstream.

Changes in sediment concentrations after the conversion of pinyon-juniper overstories to herbaceous plants on the watersheds in northern Arizona are largely insignificant (Clary et al., 1974; Baker, 1984) with one exception. Total sediments (both suspended sediments and bedloads) after the conversion of one watershed by chaining ranged up to 2.25 mg/ha annually with the largest concentration associated with a historic rainstorm with an estimated return interval of 100–150 years.

Sediment-rating curves relating suspended-sediment concentrations to streamflow discharges on watersheds in the pinyon-juniper woodlands of northern Arizona showed increased concentrations of suspended sediments because of the uprooting of trees in mechanical-conversion treatments (Lopes et al., 1999). There were no changes in the sediment-rating curve for the watershed where the chemical conversion was applied. The soil surface was not disturbed by this treatment since the herbicides were applied aerially.

Sediment concentrations did not increase following the clearcutting of oak overstories on a watershed located on the lower slopes of the Sierra Nevada (Epifanio et al., 1991).

In addition to concerns about sediment responses to tree removal, studies of chemical constituents in water samples before and after conversion of pinyon-juniper woodlands to herbaceous cover in northern Arizona found that silica (SiO_2), Ca, Mg, Na, K, bicarbonate, carbonate ($CaCO_3$), sulfate (SO_4), chloride (Cl), fluoride (F), Fe, Mn, boron (B), NO_3, and K did not exceed water-quality standards following conversion (Clary et al., 1974; Baker, 1984).

Elevated levels of NO_3 have been reported in surface runoff following the occurrence of a wildfire in the oak woodlands but none so high that they are pollution problems (Wells, 1982). Increases in the concentrations of NH_4, organic-N and Ca, Mg, Na, and K cations have also been observed after a fire.

RANGELANDS

Almost one-third of the world's land surface is classified as rangelands. These mostly arid and semiarid ecosystems include areas of shrubs with interspersed stands of grasses and forbs and communities of mostly grass species with an intermingling of forbs and occasional scattering of shrubs. The widespread occurrence of rocks and shallow soils in these ecosystems preclude land-use practices for most purposes other than grazing livestock.

Rainfall is limited on most rangelands where there is little potential for water yield improvement. Grazing in rangelands does not affect annual water yield although overgrazing can potentially increase surface runoff. Increased stormflows caused by overgrazing are usually accompanied by high rates of soil erosion and sediment transport (Branson et al., 1981).

As with woodlands, there is little potential to increase water yield through vegetative manipulations where annual precipitation is less than 450 mm/year (Hibbert, 1983). Given these constraints, watershed management of rangelands should be more concerned with managing grazing to sustain healthy vegetative cover, minimizing soil erosion and the transport of sediment and nutrients, and maintaining healthy riparian systems (see Chapter 13). Options to enhance production include rainfall harvesting on rangelands to provide water supplies for livestock and achieve better distribution of grazing animals.

Less information about the hydrologic processes of rangeland ecosystems than forests poses a constraint to implementing watershed management on rangelands (Frasier and Holland, 2004).

Vegetative Manipulations and Water Yield

Attempts at increasing water yield from rangelands have included conversion of deep-rooted shrub species to shallow-rooted herbaceous plants. Based on plot studies, Johnson et al. (1969) and Tew (1969) suggested that converting shrub communities in the Rocky Mountain Region, such as Gambel oak (*Quercus gambelii*), to shallow-rooted herbaceous plants could reduce consumptive use of water that might result in 25–75 mm of additional streamflow. This increase would be short-lived (up to 5 years) if the regrowth of shrubs is not controlled.

Increases in water yield were observed following the mechanical removal of chaparral shrubs on a 16-ha watershed in the San Dimas Experimental Forest in the San Gabriel Mountains of southern California (Dunn et al., 1988). Herbicides were applied later to control regrowth of the shrubs. The reduction in ET losses was the main factor contributing to the increase in streamflow discharges. The largest increases are observed following the occasional rainstorms in the spring and early summer while increases were not observed with larger rains when the soils were wet.

A chemical conversion of chaparral shrubs to herbaceous plants on a 100-ha watershed in central Arizona consisted of placing pelleted fenuron beneath some of the shrubs and the spraying of the other shrubs with a solution of 2,4-D and 2,4,5-T in diesel oil. While the increases in water yields following this treatment averaged only 6–12 mm annually, they represented increases up to 65% of the inherently low-streamflow discharges before the herbicidal applications (Hibbert, 1983). However, the use of such chemicals on public lands presents environmental concerns.

Streamflow discharges from watersheds in the chaparral communities of southern California and Arizona have increased temporarily as a result of wildfires that were followed by applications of herbicides to control regrowth of chaparral shrubs (Fig. 12.10). Crouse (1961) reported an increase in water yields in the first year after a wildfire on a 40.5 ha watershed on the San Dimas Experimental Forest. About 16 ha of this watershed had been treated with herbicides before the wildfire to control the growth of chaparral shrubs. Streamflow measured 2 months after the wildfire was about 275% of that expected to occur had the watershed not burned. The prefire-intermittent streamflows were converted to perennial flows following the wildfire. The loss of a protective vegetative cover, a decrease in the accumulations of litter and duff and the frequent formation of water-repellent soils following the fire were the primary mechanisms for initial increases in streamflow discharge. The regrowth of shrubs following the wildfire must be suppressed with herbicidal treatments to sustain the increases in water yields.

Water yields from watersheds up to 40 ha in the chaparral communities of central Arizona increased an average of nearly 100 mm annually following a wildfire and postfire applications of pelleted and granular fenuron and karbutylate applied by hand and aerial

FIGURE 12.10. Water yields increased following a wildfire and application of herbicides on the two watersheds on the lower elevations in a chaparral community in the southwestern USA

applications of 2,4,5-T to control the regrowth of the shrubs (Hibbert, 1983). While streamflows from these watersheds had been mostly absent in the prefire years, they became intermittent after the wildfire and remained so until the herbicides were applied at which time streamflow regimes became perennial. The results of chaparral conversions on seven watersheds in Arizona indicated that increases in annual water yield varied considerably with significant increases occurring in the higher elevations where annual precipitation exceeded 500 mm (Fig. 12.11). However, concerns about the environmental effects of applying herbicides following a wildfire have limited their use in large-scale conversion treatments at this time.

Stormflow Response to Management

Stormflow peaks from watersheds originally supporting chaparral shrubs can be higher following the mechanical conversion to herbaceous plants than on untreated watersheds because of the increases in overland flows of water following the conversion treatments (Hibbert, 1983). The occurrence of a wildfire can also create conditions that result in increases in peak stormflows. Peak flows are smaller where subsurface flows are a significant part of the streamflow regime of a treated watershed because of the increased time required for water to move through than over the soil.

Grazing livestock reduce the biomass of forage plants that in turn reduces the interception, transpiration, and soil-moisture components of the water budget for a grassland ecosystem. It is unlikely, however, that grazing intensities that are consistent with the

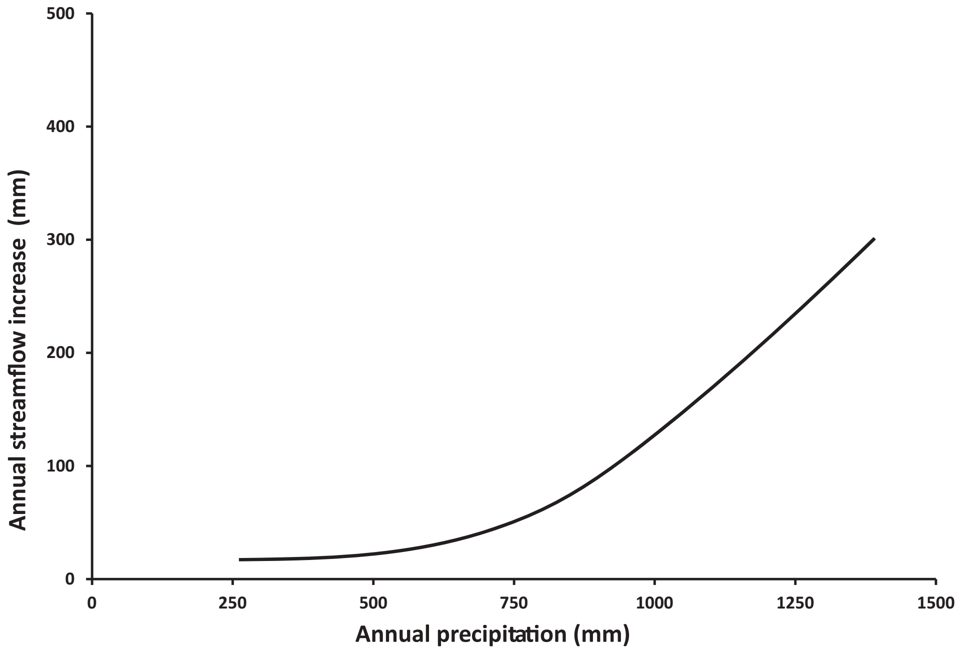

FIGURE 12.11. Annual streamflow discharges increased with increasing annual precipitation on 11 watersheds in the chaparral communities of central Arizona following wildfire and applications of herbicides to control shrub regrowth (based on data from Hibbert et al., 1982)

carrying capacity of a watershed would alter stormflow peaks or volumes (Baker, 1998). An exception could be the occasional high-peak stormflows that originate on sites where livestock grazing in excess of the carrying capacity has removed most of the protective vegetative cover.

Water Quality

The results of a few studies of the changes in the quality of streamflow water following the conversion of shrub-dominated rangelands to understories of herbaceous plants are summarized below. The quality of streams in grassland ecosystems is impacted mostly by the effects that livestock grazing have on sediment and biological characteristics.

Sediment Yields

Storage volumes in higher-order bedrock channels in the San Gabriel Mountains of southern California are generally smaller than the volumes of sediment that are produced in nonbedrock channels that have large sediment yields. It is not surprising, therefore, that Krammes (1960) reported a 10-fold increase in annual sediment yields to 55 300 kg/ha following a wildfire from a watershed in the chaparral communities of these mountains.

Sediment yields decreased from 3750 to 290 kg/ha following the conversion of semidesert shrubs to grasses that were planted after the conversion of a 44.5-ha watershed on the Walnut Gulch Experimental Watershed in southeastern Arizona (Simanton et al., 1977). Much of the reduction in sediment yields occurred once the grasses had become established.

Accumulated sediment yields in the first 3 years after a wildfire and applications of herbicides on the watersheds in the chaparral communities of central Arizona were nearly 78 m^3/ha, a value that was 25-times greater than the total sediment that had been measured in 3 prefire and 10 postfire years (Hibbert et al., 1974). These watersheds had become increasingly vulnerable to soil erosion caused by the large rainstorms shortly after the wildfire.

Chemical Characteristics

Low concentrations of fenuron were detected in the stormflows from the watershed in central Arizona where pelleted fenuron was placed beneath some of the chaparral shrubs to increase streamflow discharges (Hibbert et al., 1974). However, in excess of 760 mm of rain was recorded in the 18-month period immediately following this treatment compared to the average of about 456 mm that likely led to larger stormflows following the treatment. These larger stormflows could have "flushed" the chemical from the watershed because fenuron was not detected in the water after this period. The 2,4-D and 2,4,5-T solution that was sprayed on other shrubs was not detected.

Nutrient composition of streamflow from watersheds in the chaparral communities of the southwestern USA changed little following the occurrence of wildfire with applications of herbicides to control regrowth of the shrubs (Hibbert et al., 1974). While NO_3 concentrations increased following the initial rainstorms after these treatments, they declined to "normal" levels with the subsequent rainfall events. There were no changes in the concentrations of other nutrients.

Biological Characteristics

The biological quality of streamflow water can be altered by excessive livestock grazing as mentioned earlier in this chapter. More specially, pathogenic bacteria and protozoa in surface runoff are a concern to watershed managers because of their potentials to cause diseases. Surface water containing these pathogens must be treated for its purification

Cryptosporidium parvum, a waterborne protozoan that can be debilitating to humans (MacKenzie et al., 1994), is often carried by livestock. In situations where the linkage between livestock and *C. parvum* in surface water is known, restricting grazing in riparian corridors can minimize livestock contact with streams and other water bodies. Livestock can be provided with watering locations away from stream systems and other water bodies to control the problems of *C. parvum* in water.

UPLAND–DOWNSTREAM CONSIDERATIONS

Accommodating the interests of people living upland and downstream is one of the main tenets of IWM (Gregersen et al., 2007). Watershed managers should recognize the needs of people downstream and managers of downstream water resources must know the capabilities of upland watersheds to deliver the flows of water necessary to meet their needs. To achieve this collaboration, the respective managers of water resources should have shared knowledge of

- the volume of water yield and flow regimes of streamflow originating on upland watersheds;
- the possible losses of water flowing en route to downstream destinations in need of water supply;
- the potential for increased stormflow, sediment yields, and impairment of water quality associated with land use on upland watersheds;
- the presence and operating procedures of reservoirs; and
- the need for assessing environmental flows at regional scales.

Losses of Water En Route Downstream

Increases in water yield from upland-forested watersheds do not necessarily result in a significant increase in water yield at downstream locations where the water is needed. As the distance between upland watersheds and downstream reservoirs increases, water losses en route also increase, including bank storage and *transmission losses* within the stream channel, transpiration losses of riparian vegetation, and increased direct evaporation from the stream. Transmission losses are due to infiltration of water into the channel bottom and can diminish any water-yield increases from upstream areas, particularly in the case of ephemeral streams in dryland regions. Riparian-phreatophyte vegetation along stream courses can transpire large amounts of water.

Transmission losses can be estimated by (1) estimating the hydraulic conductivity of stream-bottom material, (2) applying the hydraulic conductivity to the total area wetted by flow, and (3) applying the above for the duration of flow. Clean-gravel and coarse-sandbed materials can have hydraulic conductivities in excess of 127 mm/h. At the other extreme, consolidated-bed material with high silt-clay content can have hydraulic conductivities of 0.03 mm/h. Transmission losses as high as 62 060 m^3/km have been reported for channels in the southwestern USA (Lane, 1983). For example, less than one-half of any increase in streamflow volumes attributed to watershed management practices or wildfire on the upland watersheds within the Verde River Basin of northern Arizona reach the reservoirs in the Phoenix metropolitan area approximately 160 km downstream (Brown and Fogel, 1987).

Water-yield improvement schemes should also take into account the evaporative losses from the reservoir pool. In arid regions, reservoir evaporation can represent a large percentage of annual streamflow at the site. van der Leeden et al. (1990) reported annual lake evaporation of 2500 mm/year in Arizona (refer to Chapter 4 for methods of estimating lake evaporation).

Another consideration is that of the timing of the increased water yield. According to most studies relating increases in water yield from forested uplands (Ice and Stednick, 2004), the largest increases in water yield usually occur during the wettest season and for the wettest years. For reservoirs to capture any increases in water yield from uplands requires that there be available space in the reservoir to store the additional water. Most reservoirs would be at or near capacity during the wettest season and the wettest years. If water flows increase when the reservoir is full, an additional water supply will be of little value.

Upstream–Downstream Flooding Relationships

To preface this discussion of changes in stormflow-flooding relationships, we need to mention that occasionally there has been confusion in dealing with the issue of flooding because of terminology problems. The technical definition of flooding, and the one used in this book, is streamflow that exceeds the bankfull capacity of a steam channel (see Chapter 6). Many people only consider a *flood* to be high flows of water leaving a channel that cause economic loss and loss of life. As we consider flooding in first-order streams to higher-order streams in downstream areas, greater damages would normally be expected when flows exceed bankfull conditions in large downstream rivers.

In most instances, it is the flooding events of greater magnitude (lower frequency of occurrence) occurring in large river systems that are the focus of flood-avoidance and flood-prevention programs. For example, the 100-year flood is used to delineate flood-prone areas for flood insurance purposes in the USA. However, in the management of first-order watersheds, being able to quantify floods of a 10-, 25-, or 50-year recurrence interval can be important in sizing culverts and bridges over small streams, or for temporary roads. Confusion often arises from interpretations of streamflow-frequency analyses and the use of recurrence interval terminology that can be misinterpreted (see Chapter 6). If a 100-year flood occurs in a given year, it does not mean that another flood of approximately the same magnitude will not occur for another 99 years. There is a probability of 1% in any given year – which increases as we consider the probability of exceeding a certain recurrence interval event over several years.

When large floods occur in downstream areas, a question often posed is to what extent did land-use activities, timber harvesting, or wildfires cause or contribute to the size of the flood? The role of forests in mitigating floods has long been controversial and led to a publication by FAO (2005) titled *Forests and floods – drowning in fiction or thriving on facts?* This publication generated considerable debate internationally – a debate that still continues in many parts of the world. As pointed out in this publication and others, the influence of forests on flooding must be considered through the lens of a hydrologist and explained in terms of hydrologic processes and scale.

Determining the impacts of upstream-peak flows on downstream flooding is difficult because of the routing and combining of the flows from different tributary channels within a watershed to downstream points of interest. Most watershed studies considering the impacts of manipulating a forest cover on stormflow events only consider the streamflows at the outlet of the (typically first order) watershed. To understand the implications of peak stormflow and flooding, the stormflow response of first-order watersheds should be considered initially.

Changes in the forest cover will more likely affect smaller floods with return periods of 5–20 years than major floods with return periods of 50 years or greater. For example, clearcutting mature aspen on a watershed in northern Minnesota increased the magnitude of annual snowmelt peaks associated with return periods up to 25 years (Fig. 12.12). Note in this figure that peak flows with the 1.5-year return period were increased 150%. The 1.5-year return-period peak flow for many stream channels is associated with bankfull conditions. Increasing the magnitude of this flow has important implications for stream-channel morphology changes.

Increases in peak flows from a first-order watershed might have little effect on downstream peaks of a river system because of the routing and desynchronization that occur. For

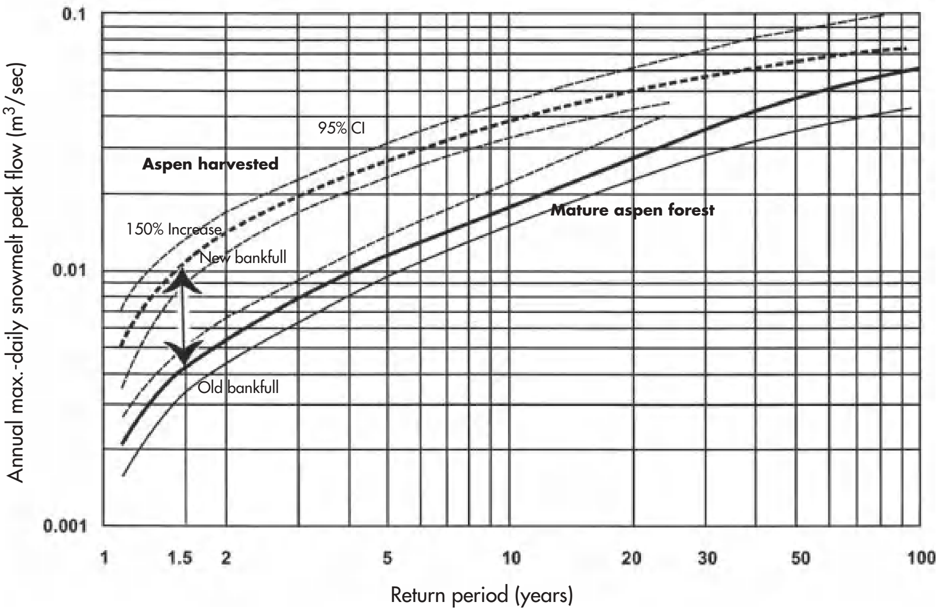

FIGURE 12.12. Changes in the magnitude and recurrence interval of annual peak flows following aspen clearcutting in northern Minnesota (from Lue (1994) as presented by Verry (2001))

example, while increased stormflow peaks originating from one tributary can increase as the result of a watershed disturbance, the magnitude of the peak discharge downstream might be reduced if the timing of upstream peak stormflows are desynchronized (Fig. 12.13). On the other hand, if stormflow volumes from these upland watersheds increase, instead of a desynchronization effect there is a cumulative and often additive effect on stormflow peaks downstream. The lesson here is that we should be more concerned about increases in the volumes of stormflow discharge than peakflow discharges in first-order watersheds.

The timing and attenuation of stormflow volumes and peaks from upland watersheds to downstream reservoirs affect reservoir operation intended to control flooding downstream of the reservoir. The use of routing procedures can help in determining how stormflow peaks and volumes from upland watersheds are attenuated or combined to result in a flood hydrograph reaching downstream locations whether reservoirs, diversion canals, and so forth. Details of streamflow-routing methods can be found in Fread (1993) and other hydrologic engineering references.

Linkages to Reservoirs

The flow of water and sediment from upland watersheds affects both the storage and operation of reservoirs that can be designed to meet one or more of the following goals:

- to provide water for agricultural irrigation, municipal uses, and hydroelectric-power generation;

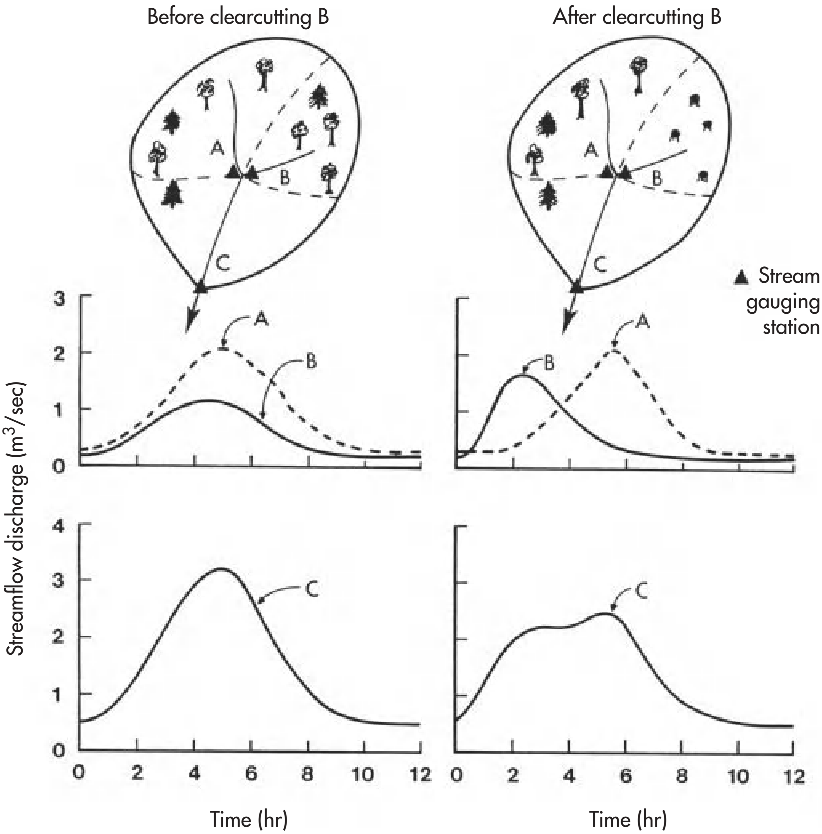

FIGURE 12.13. Effects of forest removal on upstream and downstream stormflow hydrographs where desynchronization of stormflow hydrographs occur

- to maintain instream flows during the dry season; and
- to provide storage to mitigate flood hazards.

Whether a reservoir remains operational to meet these goals depends to a large extent on the management and condition of the upland watersheds contributing water to the reservoir. For example, watershed management practices that maintain high rates of infiltration and low flows of surface runoff should be considered when flood control is an objective. Wetland areas within a watershed attenuate stormflow peaks, and, therefore, their protection should also be part of the overall watershed management program when flooding is a concern. However, an accelerated rate of sedimentation resulting from land use on upland watersheds can cause a loss of reservoir storage and the ability of that reservoir to meet the demands for which it was designed.

Storage Requirements

Determining the storage capacity of a reservoir requires that the inflows of water to the reservoir and rates of sediment deposition be balanced against projected demands for water. The total storage capacity of a reservoir is the sum of the active storage and dead storage.

Active storage is that needed to meet all demands; that is, to prevent shortages and (in some cases) to provide flood-control benefits. *Dead storage* is that part of the reservoir allocated to trap and hold sediment and, therefore, must be of sufficient volume to prevent sedimentation from affecting the active storage for a period equal to the economic-design life of the reservoir. The gates in a dam to release water from the reservoir are located above the upper level of the dead-storage zone.

General Operating Procedures

The general operating procedures of a reservoir can best be understood by considering the storage requirements for the multipurpose reservoir illustrated in Figure 12.14. This reservoir has storage space allocated for both conservation (water supply) and flood-control purposes. Within the conservation storage is a buffer zone, the top of which is used as a threshold to allocate water releases from the reservoir during critical dry periods. Once the elevation of the reservoir pool drops to the top of the buffer zone, water can be released only for those purposes determined to be most important such as providing municipal-water supplies. Lesser needs will not be met until the pool elevation is higher than the top of the buffer pool.

Several buffer zones can be used as a mechanism to allocate water on a priority basis in periods of shortages. If the reservoir has a hydroelectric-power-generating capacity, the amount of head required to drive the turbines and the corresponding storage would be an added operational zone in the reservoir. Once sediment begins to encroach into the respective storage spaces, operating the reservoir to meet the respective demands becomes more difficult. As shown in Figure 12.14, sediment does not necessarily settle out only in the dead-storage space. Coarse materials are deposited at inflow points in the upper reaches of the reservoir pool while finer sediments tend to settle out near the dam.

The capability of a reservoir to meet all of the anticipated demands for water becomes limited as sediment fills the active-storage space. Storage volumes in the conservation pool can be inadequate to meet demands in periods of drought or there might not be adequate storage space in the flood-control pool to control major flood events. Some of these problems

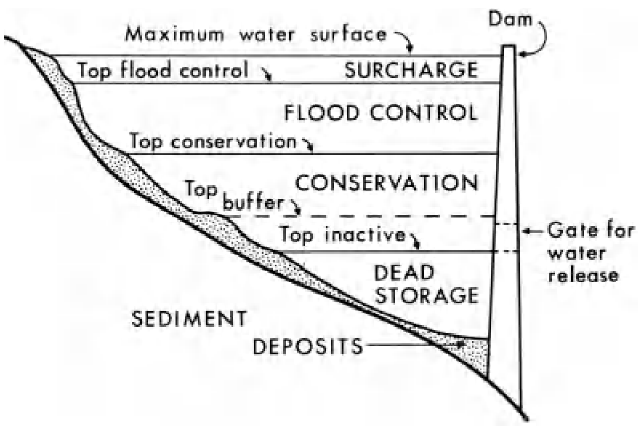

FIGURE 12.14. Storage allocation for a multipurpose reservoir with associated sedimentation

can be addressed by applying an *operational rule curve approach* to reallocate storage space for different purposes seasonally. Rule curves provide operational guidelines for the most efficient use of reservoir storage. For example, the season in which irrigation supplies are needed is often the period in which the threat of large floods is minimal. Therefore, the elevation of the conservation pool can be raised into the normal flood-control pool during that time. When the flood season approaches, the conservation pool is then lowered to provide flood-control space. Rule curves allow demands to be met with a smaller total storage capacity in the reservoir. But even rule-curve operations might not suffice when sedimentation becomes excessive.

Cumulative Watershed Effects

The cumulative watershed effects (CWEs) of changing watershed conditions caused by planned and unplanned disturbances present a challenge to watershed managers and users of the natural resources on wildland watersheds. Importantly, watersheds should be managed such that the management practices and other land-use activities do not cumulatively have adverse impacts on the other beneficial uses. Obtaining the multiple benefits available on many wildland ecosystems is often compatible with efforts to maintain or increase the flows of high-quality water. For example, removal of chaparral shrubs to increase forage production can also increase water yield in some instances and can benefit certain wildlife species and reduce fire hazards. However, the same reductions in shrubs can be detrimental to other wildlife species. A watershed manager must understand how the collective set of management and land-use actions on a watershed complement or compete with multiple-resource objectives and mitigate unwanted effects. The overall objective is to achieve production goals without adversely affecting soils and water flows.

People gain a better understanding of the spatial and temporal relationships of watershed management practices and other land-use activities and their effects on the production of water, sediment, chemicals, and the other watershed values by recognizing the CWEs of these practices and activities. Many CWEs can be determined by understanding the hydrologic, water-quality, and geomorphic effects of various land uses and natural processes across a watershed. Many of these relationships are known and the challenge is to understand the interactions that occur – where effects are compounded and where they are diminished. An interdisciplinary approach incorporating hydrology, geomorphology, and ecology into watershed management and other land-use activities is needed to understand and appreciate the impacts of CWEs on water yields, other streamflow characteristics, and water quality.

Water Yield and Streamflow Regimen

The amount and timing of water flow from first-order watersheds and the downstream flow and sediment relationships in the channel are all affected by environmental changes, watershed management practices, and land-use activities on the watershed. Changes in vegetative composition, density, age structures, and continuity across the landscape can affect ET losses and, as a consequence, influence water yields and timing streamflow discharge at different parts of a watershed. Depending on the extent that a watershed is affected by these changes and the antecedent soil-moisture conditions at the time of the changes, these modifications can also increase stormflow volumes and peaks. However, alterations of streamflow regimes can be byproducts of vegetative management practices implemented for other purposes such as harvesting timber, increasing herbage (forage) production, or enhancing wildlife habitats.

Trampling of a watershed by excessive numbers of livestock compacts the soil surface and, in doing so, alter its hydrologic properties. For example, overgrazing can compact the soil, reduce infiltration capacities, and increase surface runoff and soil erosion. Such increases in runoff from one part of watershed can be additive to increases in runoff from recreational developments, roads, and the fragmentation of forest cover in another location. Unsurfaced roads are often compacted and, as a consequence, generate increases in overland flows of water, and ultimately higher streamflow volumes and sediment delivery downstream. Roads and trails are possible sources of increased soil erosion because of the exposure to erodible soil and subsoil in their construction can reduce infiltration of water on the corridors surfaces, increased gradients on cutslopes and fillslopes, and concentration of overland flow from precipitation excess and interception of subsurface flow. Downstream flows and sediment loads reflect these CWEs.

Wildland watersheds are often impacted by large wildfires with the hydrologic functioning of the watersheds altered as a consequence. If other parts of a watershed have undergone timber harvesting, and yet others the development of home sites, there can be increases in stormflow in excess of what would be anticipated from any one of these activities. For example, the burned areas can result in water-repellent (hydrophobic) layers in soil and increased surface runoff. Flood peaks can also be increased, low-flow discharges lowered, and soil moisture decreased. In addition, lowered infiltration capacities and percolation to groundwater aquifers can lead to reduced levels of flow during the dry season.

Water Quality

Land-use activities that change the volume or timing of streamflow can also affect the rates of sediment and chemical transport from a watershed because they are dependent on the magnitude and rate of water flows. Timber harvesting, silvicultural cuttings, or wildfire can cause a short-term increase in the concentrations of sediments and chemicals in streams. For example, increases in N concentrations can result from reduced N uptake, increased subsurface flow, increased alteration of N to leachable forms, or increased volumes of decaying organisms. Other management activities that contribute to these effects include herbicidal applications or prescribed burning.

Some vegetative manipulations and land-use activities alter chemical pollutants by causing them to interact with the environment in different ways. For example, limestone road gravels can moderate the pH of surface runoff generated by acid rainfall. Acid rainfall often increases the mobility of nutrients and, therefore, contributes to nutrient deficiencies. Aluminum released by acid rainfall is less toxic to fish populations when streams are enriched in silica. Vegetative conversions can also change the spatial distribution of the resulting nitrogen mineralization because the plants lend different mineralization potentials to the soil. Chemicals reaching stream channels can be redeposited in downstream reservoirs where dissolved pollutants concentrate if evaporation rates are high.

Assessing Environmental Flows

The provision of “ecosystem services” through IWM would include a myriad of environmental benefits attributed to healthy watersheds, including low rates of soil erosion, naturally variable flows of high-quality water, provision of upland and aquatic habitat, maintaining stable stream channels, and so forth. The linkage between the quantity, timing, and quality of streamflow from upland watersheds that are required to sustain downstream aquatic ecosystems and well-being of humans who are dependent on those ecosystems are called "environmental flows" as described by Poff et al. (2010). They point out the need for a framework to assess environmental flows for streams and rivers at a regional scale. This framework, the ecological limits of hydrologic alteration, is a synthesis of a number of existing hydrologic techniques and environmental-flow methods that are currently being used to various degrees with the goal of supporting comprehensive regional-flow management (Poff et al., 2010). These authors state that

> The flow regime is a primary determinant of the structure and function of aquatic and riparian ecosystems for streams and rivers. Hydrologic alteration has impaired riverine ecosystems on a global scale, and the pace and intensity of human development greatly exceeds the ability of scientists to assess the effects on a river-by-river basis. Current scientific understanding of hydrologic controls on riverine ecosystems and experience gained from individual river studies support development of environmental flow standards at the regional scale.

The development of environmental-flow standards is part of a movement to recognize the value of natural capital or ecosystem services; wildland watersheds have ecological limits that can be exceeded by poor management decisions.

Summary and Learning Points

Annual water yield, stormflow peaks and volumes, and low flows can be affected by activities on upland watersheds including timber harvesting, wildfires, roads, and vegetative conversions. Some key points that should be understood concerning the above relationships include

1. What changes in vegetative cover usually result in an increase in the quantity of water yield? What are the exceptions?
2. What methods are available to estimate changes in water yield caused by changes in vegetative cover?
3. How can land-use practices affect streamflow during periods of high flows and of low flows?
4. What are some examples of changes in a watershed that can result in
   - higher flows during the dry season;
   - higher flows during the wet season; and
   - lower flows during both the dry and the wet season?
5. Does clearcutting of forests cause flooding to increase? If so, explain why.
6. How do wildfire and prescribed fire differ in their effects on streamflow regimes?
7. In what way do land-use activities and environmental change affect hydrologic processes on watersheds and the ultimate streamflow response? What are the relations between hydrologic processes on a watershed and CWEs?
8. What factors are important in determining how much of a change in water yields from an upland watershed might reach downstream reservoirs?

The protection and maintenance of high-quality surface water are also fundamental goals of watershed management. This chapter considers issues of importance to managers of the quality of water flowing from upland watersheds and describes some of the impacts of land-use activities and disturbances on these characteristics and the aquatic ecosystem. In this regard you should be able to

1. Discuss the more important issues confronting managers concerned about maintaining flows of high-quality water from upland watersheds.
2. Discuss how tree-cutting activities, wildfire, and roads can affect water quality.
3. Describe the general management measures needed to mitigate or prevent water-quality problems.
4. Describe the types of water-quality monitoring programs and discuss how to locate sampling stations for each.

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CHAPTER 13 Managing Riparian Communities and Wetlands

INTRODUCTION

Special attention is given to the functioning and management of riparian communities and wetlands in this chapter. Riparian areas have been degraded and wetlands drained with often minimal recognition of the cumulative effects of these actions in many parts of the United States and elsewhere in the world. However, their values are currently recognized in terms of their hydrologic functions, impacts on water quality, and ecological importance. Riparian communities and wetlands are integral components of watersheds that influence aquatic habitats and the flow of water, sediment, and other pollutants in streams and rivers. Consequently, when riparian areas and wetlands are degraded or destroyed there can be serious impacts on the hydrologic response and water quality of watersheds. Actions taken to restore or rehabilitate riparian areas and wetlands can be important assets in improving hydrologic function and mitigating impaired water quality of watersheds; their importance to integrated watershed management (IWM) of wildland, agricultural, and urban watersheds should become evident after reading this chapter.

RIPARIAN COMMUNITIES

Riparian communities occupy ecological sites situated between terrestrial landscapes and aquatic ecosystems. The vegetation in riparian areas is characterized by plants that require the regular presence of free or unbound water flowing through them (Fig. 13.1). Surface water and groundwater usually come together in riparian zones; important exchanges of water, nutrients, and minerals between groundwater and stream water occur adjacent to the channel and in the *hyporheic zone* that is the zone of saturated sediments below the streambed (discussed in Chapter 7). While these streamside ecosystems occupy only a

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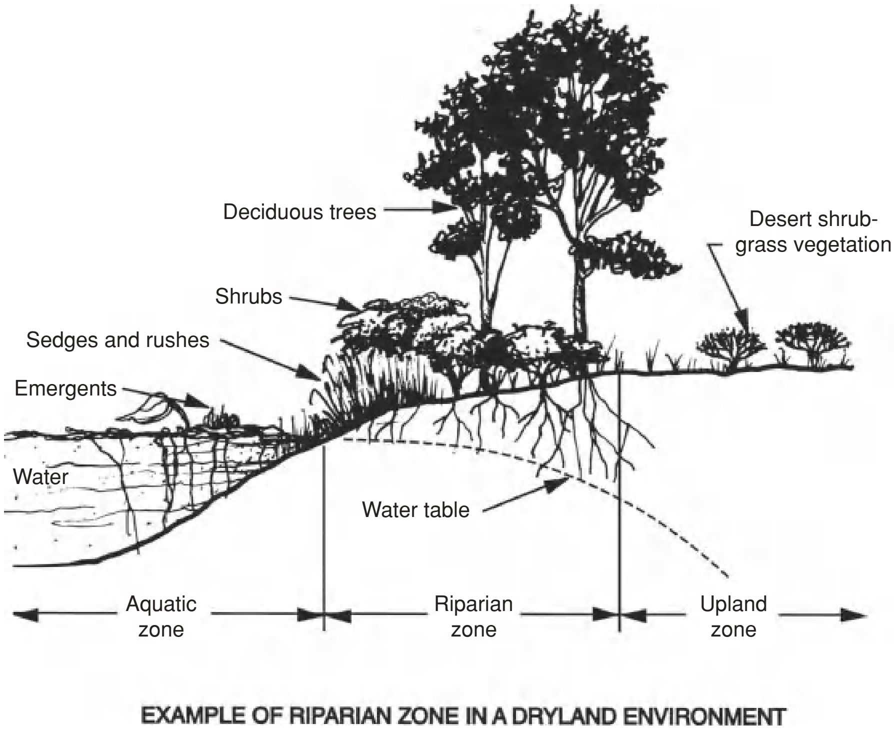


FIGURE 13.1. Riparian ecosystems are most often associated with the presence of vegetation that requires large amounts of free or unbound water. The ecosystem illustrated in this figure is that in a dryland environment (from Thomas et al., 1979)

small part of the area of a watershed, they have disproportionate values from a hydrologic and ecological standpoint.

Riparian vegetative communities are often unique from those in the adjacent uplands and provide many important roles. They affect the microclimate by providing shade to streams, resulting in lower stream temperatures that are more favorable for fish habitat. These communities also provide a critical habitat for many invertebrates, amphibians, reptiles, small and large mammals, and birds. Vegetation acts as a filter, or buffer, because it traps the overland movement of sediments and can take up nutrients from the saturated and unsaturated zones adjacent to water bodies. This vegetation also plays a role in supplying habitat for microorganisms and organic matter to the soil and stream. Litterfall, when covered with sediment, decomposes quickly and releases both nutrients and humus to the soil. Structurally intact living and dead plant material is colonized by microorganisms, providing biochemical transformations and food for macroinvertebrates which serve as food for fish and wildlife species.

Riparian vegetation consists of trees, shrubs, grasses, rushes, and sedges in varying combinations and is adapted to well-drained but periodically flooded soils. Stream riparian systems typically have transported coarse soils and are adapted to high-magnitude, short-duration flood events. The physical structure of woody-riparian vegetation consisting of tall, deep-rooted trees and shrubs enables them to survive flooding. Riparian trees provide large woody materials that can improve the stability of stream channels. Riparian vegetation, and particularly riparian forests, exhibits high rates of *ET* that are usually near *PET* rates (Tabacchi et al., 2000). Along streambanks, persistent high *ET* rates result in reduced pore pressures which when combined with the root strength of woody vegetation lead to greater streambank stability. Extensive root systems and organic matter help to bind the soil and promote high infiltration rates. Trees and associated vegetation along stream channels can trap sediments during flood events leading to floodplain succession and ultimately community succession.

Riparian corridors function as buffers that protect stream systems or lakes from disturbances on the surrounding uplands. They also provide habitats for a variety of wildlife species, are often grazed by livestock, and are sites for picnicking, camping, and fishing. Therefore, maintaining the integrity of riparian systems and avoiding degradation by excessive tree-cutting activities, livestock grazing, road crossings, and recreational pressures are crucial.

Recognizing the characteristics of a healthy and well-functioning riparian community and understanding the linkages between a riparian community and its watershed provide a basis for formulating management practices to sustain riparian ecosystems for their many benefits. These characteristics and linkages are summarized below as a prelude to discussing the management practices implemented to maintain these conditions.

Characteristics of a Healthy Riparian Community

A *healthy riparian community* along a stream channel is characterized by a dynamic equilibrium between stream channel deposition (aggradation) and downcutting (degradation) by erosion of the embedded stream channel. Such an ecosystem is able to maintain this equilibrium between the streamflow forces acting to produce change and the resistance of vegetative, geomorphic, and structural features to the change (Fig. 13.2; see discussion in Chapter 9). Over time this riparian equilibrium is sufficiently stable and dynamic to compensate for external disturbances from uplands (Verry et al., 2000; Baker et al., 2004). Resistance to a change is expressed as resiliency and results largely from adjustments among factors including vegetation, channel depth, and stream morphology operating simultaneously within the riparian system to convey the flow of water and sediment without significantly altering the equilibrium.

Short-term increases in surface runoff from uplands of the riparian community caused by disturbances such as timber harvesting or wildfire can increase the velocity and volume of streamflow, increase surface soil erosion, and increase erosion of the streambank and the channel itself, thereby increasing the deposition of sediment in the channel and floodplain when flows recede. Under these conditions, a healthy riparian ecosystem will oscillate back and forth so that only a little change is likely to occur in its dynamic equilibrium (LaFayette and DeBano, 1990; Baker et al., 1998). Furthermore, the equilibrium condition can be re-established following a disturbance when the resiliency of the linked riparian–watershed system has not been altered significantly by the disturbance. However, a healthy riparian ecosystem can be adversely affected when degradation or aggradation overwhelms this equilibrium. Degradation processes will result in *unhealthy riparian conditions* more rapidly than processes of aggradation due to land loss.

Wildfires that burn the upland watershed, the riparian corridor, or both can adversely affect a healthy riparian ecosystem. When only the upland watershed has been burned,

FIGURE 13.2. A healthy riparian community is able to maintain an equilibrium between the streamflow forces acting to produce change and the resistance of vegetative, geomorphic, and structural features to the change. Riparian area in Honduras (Photograph by Mark Davidson) (For a color version of this photo, see the color plate section)

a riparian corridor can act as a buffer between the burned watershed and the stream system. This buffer can filter out eroded soil particles and other pollutants entrained in the often increased surface runoff occurring after a fire. However, undesirable impacts can be exacerbated when both the riparian community and the upland watershed are severely burned by a wildfire (see below).

Riparian–Watershed Linkages

Understanding the importance of riparian functions and health to the overall condition of a watershed (Box 13.1) is a key to managing riparian ecosystems in a balanced, environmentally sound, and holistic manner. The nature of the linkages between riparian and

Box 13.1

Characteristics of Watershed Condition

Watershed condition is a term that reflects the status (health) of a watershed in relation to its hydrologic function and soil productivity. *Hydrologic function* in this case refers to the ability of the watershed to receive and process precipitation into streamflow. *Soil productivity* reflects the capabilities of the watershed for sustaining plant growth and plant communities or the natural sequences of plant communities.

On a watershed in good condition:

- Precipitation infiltrates into the soil.
- Precipitation does not contribute excessively to soil erosion since the resultant overland flow of water does not dislodge and then move soil particles.
- Soil productivity is sustained.
- Streamflow response to precipitation is relatively slow relative to the underlying geology.
- Baseflow (often groundwater flow) of perennial streams is sustained between storms.

While on a watershed in poor condition:

- A relatively large portion of the precipitation input flows over the soil surface due to reduced infiltration capacity.
- Excessive soil erosion occurs during precipitation events and soil productivity diminishes.
- Streamflow response to precipitation is rapid; there is little or no storage in the watershed.
- There is little or no baseflow between storms in streams that were once perennial.

watershed health can be illustrated by initially considering historical accounts of the hydrologic functioning of the respective systems. Before large-scale interventions by people, riparian corridors were generally stable in aggregating stream networks containing large amounts of woody debris (see below) and supporting abundant aquatic biota (Cooperrider and Hendricks, 1937; Leopold, 1951; Minckley and Rinne, 1985). Under these conditions, headwater streams supplied continuing flows of small- and large-woody debris and accumulations of logs and other woody debris that formed log steps in small streams with larger accumulations of these materials along and within higher order streams.

Floodplains, channel structures, and the associated vegetation dissipate stream energy, control sediment transport, and regulate water flows to provide the downstream environments that maintain healthy riparian communities. The dissipation of stream energy also increases the percolation of water into subsurface storage. That portion occurring as detention storage of subsurface water in high-elevation soils and aquifers, flows slowly into riparian areas and can sustain "late-season" low flows that benefit downstream riparian areas.

Improper management or unwise uses of riparian corridors and upland watersheds upsets the historical linkages between riparian health and watershed condition. A degraded or unhealthy riparian area is often present when its watershed is in poor condition (Heede, 1986; Baker et al., 1998). In contrast, a healthy riparian ecosystem is likely to be found within a watershed that is in good condition. Sometimes the linkages between riparian and watershed conditions are less apparent because of the lag in time between the changes that are occurring on upland areas of a watershed and the effects of these changes on downstream riparian communities.

Riparian Management

Increasing water for downstream users by eradicating riparian plants that consume high amounts of shallow groundwater in floodplain and streamside areas was an early focus of riparian management in the southwestern United States (Horton and Campbell, 1974; Johnson et al., 1985). However, pressure from environmental groups and the general public to provide more holistic management of riparian communities largely halted this practice by the early 1980s. Managing riparian communities throughout the United States currently centers on achieving one or more of the following goals:

- sustaining protection of streams from disturbances on watersheds by maintaining buffer strips of streamside vegetation;
- retaining or establishing woody debris in riparian corridors and upland watersheds to enhance the stability of soils and stream channels;
- controlling wildfires to minimize their impacts;
- implementing rehabilitation activities when necessary to restore the functioning of degraded riparian ecosystems; and
- meeting instream flow requirements.

Sustaining Buffer Strips

Maintaining the integrity of riparian buffer strips is an effective way of reducing the velocity of overland flows of water from the surrounding uplands and, in doing so, depositing some of the entrained soil particles and other pollutants before reaching a stream channel (Neary et al., 2010). The processes involved in the filtering of sediment- and pollutant-laden surface runoff by a buffer strip are illustrated in Figure 13.3. The effectiveness of a buffer strip in maintaining water quality is dependent on its vegetative characteristics and its width, the velocity of the water flow, the size of eroded soil particles, and the topography of the site. Buffer strips are particularly effective in serving as a filter for sediment in overland flow occurring as rills and interrill flow from uplands. They are not effective when there are continuous gullies from uplands that cut through the riparian community and discharge directly into the stream channel. Similarly, when culverts or drainpipes transport water and pollutants from uplands through a riparian zone, there are no opportunities for the riparian vegetation and soil system to reduce flows and loading to the stream.

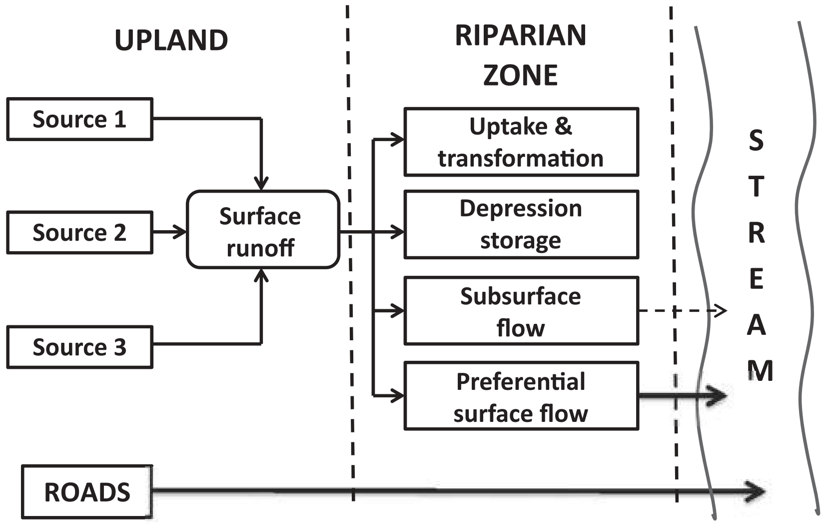


FIGURE 13.3. A schematic of how dissolved constituents and entrained soil particles move through a riparian buffer strip (RBS) (modified from Comerford et al., 1992).

The benefits of riparian buffer strips left intact or established where they are lacking include:

- protecting stream channels and the edges of other water bodies from upland perturbations (tree cutting, livestock grazing, wildfire, construction, agricultural cultivation, etc.);
- protecting and sustaining habitats for fish and wildlife and maintaining scenic qualities;
- providing large-woody materials to maintain or improve the hydrologic functioning of the stream system and other water bodies; and
- reducing solar radiation impinging on the surface of streams to provide cooler water favoring higher-quality aquatic organisms while maintaining higher levels of dissolved oxygen – a major function of riparian vegetation along small headwater streams.

Guidelines for establishing riparian buffer strips are outlined in Table 13.1. However, Corner and Bassman (1993) suggest that the width of a buffer strip should be determined as a function of parameters such as the slope, soil permeability, soil erodibility, and the nature of the management of upland areas rather than the specified widths presented in Table 13.1. Ultimately, the effectiveness of a buffer strip in protecting a stream from nonpoint pollution is related largely to the management of upland areas.

General guidelines for managing riparian buffer strips once they are established are illustrated in Figure 13.4. Additionally, structural approaches might become necessary to control gully erosion because buffer strips are not effective in controlling the channelized flows of water originating on the upslope of buffer strips. Typically, gully erosion suggests a significant change in either contributing drainage area or land-use management; this issue must be addressed above the riparian zone.

TABLE 13.1. Guidelines for establishing buffer strips in the western United States

		Buffer strip requirements		
State	**Stream class**	**Width**	**Shade or canopy**	**Leave trees**
Idaho	Class I[a]	Fixed minimum (75 ft)	75% current shade[f]	Yes; #/1000 ft dependent on stream width[h]
	Class II[b]	Fixed minimum (5 ft)	None	None
Washington	Type 1, 2, and 3[a]	Variable by stream width (5–100 ft)[c]	50%; 75% if temperature $>60°F$	Yes; #/1000 ft dependent on stream width and bed material
	Type 4[b]	None	None	25/1000 ft $>$ 6 in. diameter
California	Class I and Class II[a]	Variable by slope and stream class (50–200 ft)	50% overstory and/or understory; dependent on slope and stream class	Yes; # to be determined by canopy density
	Class III[b]	None[d]	50% understory[g]	None[g]
Oregon	Class I[a]	Variable; three times stream width (25–100 ft)	50% existing canopy, 75% existing shade	Yes; #/1000 ft and basal area/1000 ft by stream width
	Class II special protection[b]	None[e]	75% existing shade	None

Source: Belt et al. (1992).
[a]Human water supply or fisheries use.
[b]Streams capable of sediment transport (California) or other influence (Idaho and Washington) or significant impact (Oregon) on downstream waters.
[c]May range as high as 300 ft for some types of timber harvest.
[d]To be determined by field inspection.
[e]In eastern Oregon, operators are required to "leave stabilization strips of undergrowth ... sufficient to prevent washing of sediment into Class I streams below."
[f]In Idaho, the shade requirement is specifically designed to maintain stream temperatures.
[g]Residual vegetation must be sufficient to prevent degradation of downstream beneficial uses.
[h]In Idaho, the leave tree requirement is specifically designed to provide for the recruitment of large-organic debris.

Retaining or Establishing Woody Debris

The presence of *large-woody debris*, *large-organic debris*, and *coarse-woody debris* within a riparian corridor and its watershed can increase the stability of a stream channel (Box 13.2). Large-woody debris and large-organic debris can improve aquatic habitats by reducing the velocity of water flow and providing refugia for invertebrates and fish (Dolph et al., 2012). Large, affixed logs extending partially across a stream channel can deflect streamflow laterally to cause the flow of water to widen across the streambed. Even when a stream channel becomes so wide that the logs cannot span the channel, the debris accumulations along the banks can cause meandering cutoffs and create secondary stream-channel systems or trap sediment and reduce channel width/depth ratio. Large-woody debris and large-organic debris also create variability in stream channel depth by creating scour pools

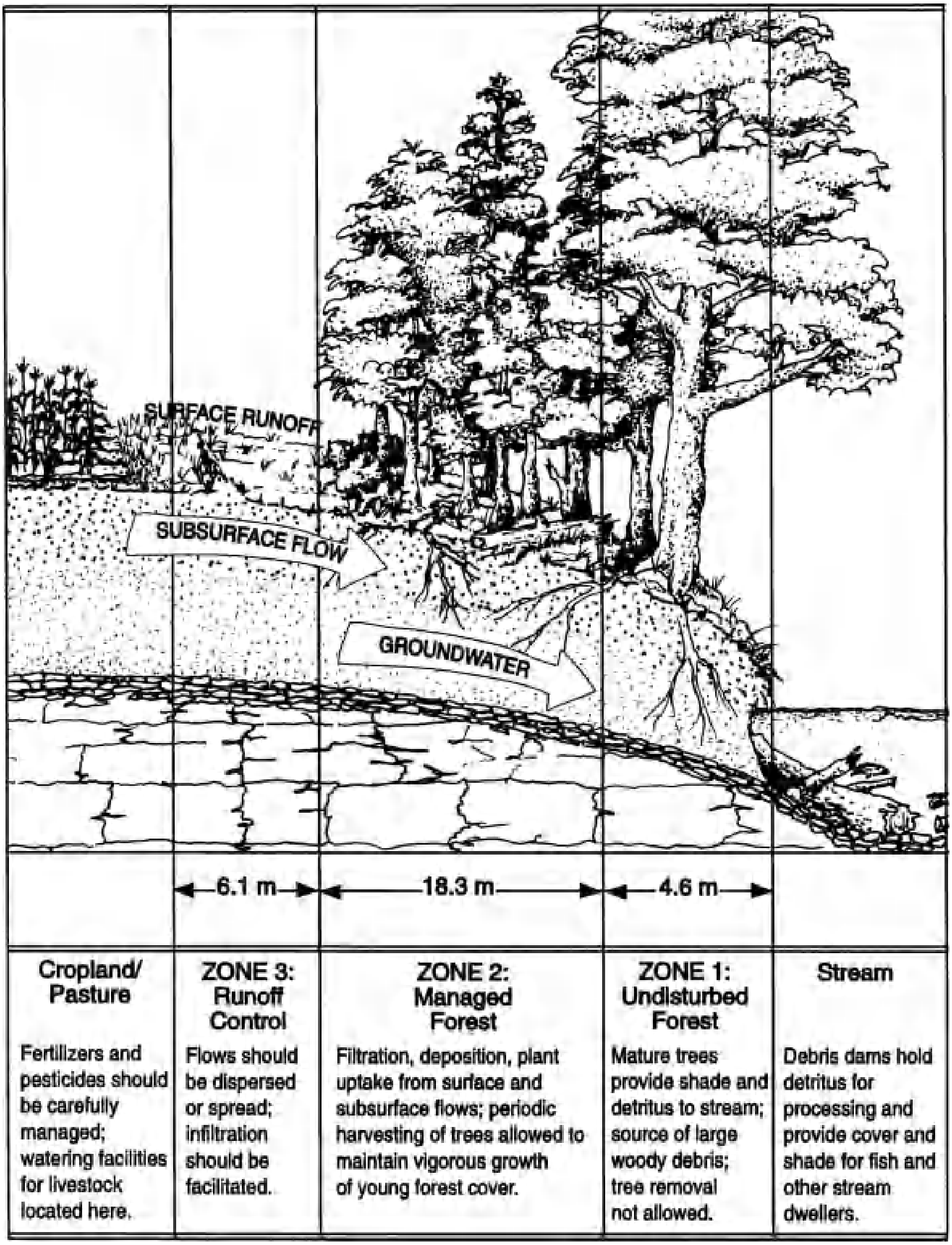


FIGURE 13.4. General guidelines for the management of established riparian buffer strips (from Welsch, 1991)

Box 13.2

Definitions and Origins of Woody Debris

Large-woody debris is any piece of relatively stable woody material at least 1 m in length and 10 cm in diameter (Dolloff, 1996). Large-organic debris consists of large woody materials that intrude into a stream channel, including large tree stems and root wads (American Fisheries Society 1985). Large-woody debris and large-organic debris are found within the riparian corridor while coarse-woody debris is found on the surrounding watershed. Large-woody debris and large-organic debris are created when trees die and fall to the ground that is near a riparian corridor. Neither large-woody debris nor large-organic debris is transported long distances from where they fell except in excessive surface-runoff events. Bank undercutting of streamside trees over time will add large-organic debris to stream channels. Downstream movement of this debris is related to the length of individual pieces (Lienkaemper and Swanson, 1987). Most of the pieces of debris that are moved are shorter than the bankfull width of the stream. Those not moved can trap sediment and raise the bed height until the wood decomposes or is rearranged by a large stormflow event.

downstream from obstructions. As a result, this debris helps to maintain a diverse riparian ecosystem and aquatic habitat by:

- anchoring and stabilizing the position of the pools in the direction of the streamflow;
- creating riparian connectivity via backwater along stream margin and forming secondary channel systems for fish spawning; and
- increasing or decreasing stream width-to-depth variability which creates refugia habitat.

Coarse-woody debris improves soil properties by enhancing the soil structure and stabilizing the soil surface while moderating the responses of a watershed to precipitation events. Vegetation, litter, and other organic material comprising the coarse-woody debris provide a soil surface capable of absorbing the energy of precipitation, increasing infiltration, and slowly transferring subsurface water flow downslope. The improved soil structure allows the soil to store and transmit water slowly from the soil surface to underground storage or flow, thereby providing temporary storage of water throughout the watershed. The slow subsurface flow of water also decreases the erosive potential produced by overland flows, especially on steep slopes.

Controlling Wildfires

Wildfires occurring in both riparian communities and their watersheds can have devastating effects when they are severe. Fire rarely occurs only in the riparian corridor unless a controlled burn is prescribed for this purpose. If the protective riparian buffer strips between the hillslopes and stream systems are lost or severely damaged by a wildfire, there can be considerable disruption to the water flow and sediment equilibrium that characterizes a healthy riparian community. Debris flows and small landslides from hillslopes are frequent on steep and erodible topography after a wildfire. The severity of these impacts is related mostly to the intensity of the fire (DeBano and Neary, 1996). A high-intensity wildfire can consume all or nearly all of the riparian vegetation while a low-intensity burn is likely to have a less severe impact on the riparian plant community. The rate of the recovery of a riparian ecosystem following a wildfire is determined largely by the combined disturbance of both the burn and the postfire streamflow regime.

Minimizing the effects of a wildfire on riparian communities requires timely and effective suppression. Monitoring and evaluation of postfire effects are necessary to determine the kind of rehabilitation activities that will help return the burned riparian corridor to a functioning status approaching that close to the prefire condition as quickly as possible. In some instances, the prefire condition might not be realistically attainable in the near future.

Rehabilitation Activities

Riparian systems and their plant communities can suffer a loss of functioning even when the "best management" practices (BMPs) are followed (discussed in Chapter 14). Rehabilitation of riparian systems can become necessary in such instances. To rehabilitate a riparian community requires knowledge of the linkages among the riparian community, its watershed, and the stream channel. Unfortunately, there are no standardized rehabilitation guidelines for riparian systems because of the inherent variability from one riparian site to another. Rehabilitation activities such as adding or removing vegetation, constructing riffle formations, building bank protection, or constructing check dams that have proven to be successful in one situation can be a failure when applied elsewhere. Therefore, an *adaptive management approach* to rehabilitate riparian systems requires:

- having an understanding of the basic hydrologic processes involved;
- knowing as much as possible about the history of the riparian area being treated; and
- making perceptive observations and using appropriate expertise to develop viable options for rehabilitation efforts.

Roads and trails on steep slopes or erodible soils in close proximity to a riparian community or a stream crossing can degrade the quality of water flowing in the stream (Mattson et al., 2000). As a consequence, rehabilitation often includes restricting the construction of roads, trails, or other corridors near to, or within, the riparian corridor or decommissioning them if they are already in place.

A host of references concerning the rehabilitation of riparian communities and wetland ecosystems have been summarized by Schneller-McDonald et al. (1990). Included in these references is information on:

- planning, establishing goals and objectives, and designing general techniques of rehabilitation;
- specific techniques of rehabilitation including planting, fencing, land forming, installing instream devices, and soil treatments; and
- monitoring and evaluating the effectiveness of these rehabilitation efforts.

Meeting Instream Flow Requirements

A watershed manager should also be aware of the need to meet instream flow requirements in situations where the streamflow has already been fully committed to previous uses. *Instream flow* is the streamflow required to satisfy the conjunctive demands on water while it is still in a channel (American Fisheries Society, 1985). *Instream flow requirements* are streamflow-discharge rates that sustain a predetermined streamflow volume. *Instream flow rights* are legal rights (entitlements) to use surface water within a specified reach of a stream channel for its designate use or combination of uses. The use or combination of uses must be nonconsumptive with exception of the "normal needs" of aquatic vegetation, wildlife, and fish and other aquatic organisms. An instream flow right protects a designated flow of water passing through the reach of a stream from depletion by newer users of the water, and, therefore, it is important when upstream developments, diversions, or transfers could threaten existing streamflow regimes. Another benefit of an instream flow right is the protection afforded to the diversity of riparian plant and animal species that live in or along the water.

Other Considerations

Knowledge of the stream continuum model, the concept of river health, and the influence of riparian vegetation on flooding are needed in planning and implementing a strategy for managing riparian communities.

The Stream Continuum Model. The stream continuum model developed by Vannote et al. (1980) provides an ecological rationale for maintaining a riparian corridor along a stream system or other water bodies. This conceptual model recognizes a watershed as a landscape that extends from the smallest headwater streams to the mouth of a large river. The physical and biological variables represented on the watershed represent a gradient in which a continuum of biological, physical, and geomorphic adjustments occurs. A variety of plant and animal species and environmental processes are related to geomorphic-channel processes, climate shifts, flood regimes, and the influences on the upstream-fluvial corridor. The continuously changing environment of a stream system that is depicted by the stream continuum model results in a variety of life cycles of organisms and a diversity of biogeochemical cycles and rates as organisms adapt to disturbance regimes over broad spatial and temporal scales.

More recently the continuum model has been expanded to acknowledge repeating cross-sectional linkages or the features found from the channel through the valley. The focus is on the transverse biocomplexity related to the longitudinal changes defined by Vannote et al. (1980). Thorp et al. (2006) developed the Riverine Ecosystem Synthesis which highlights hydrogeomorphic patches that provide riparian connectivity. The concept of riparian connectivity has become increasingly important to the holistic approach of IWM.

The Concept of River Health. The concept of *river health* can be useful in planning for the management of riparian communities on a river-basin scale (Boulton, 1999). River health is similar to the notion of ecosystem health in that both ecological and human values are incorporated into a philosophical framework. The ecological values consist of ecological integrity that includes the capacity to maintain natural, balanced, integrative, adaptive biological systems and the resilience to stress (such as the ability to recover following disturbance). The human values include economic goods (water for irrigation and industry, clean water for domestic use, and an environment for recreation and spiritual renewal) and services (cleansing of water, producing fish and shellfish, and providing aesthetic values). The indicators that might be used in assessing river health vary from one system to another.

Influences of Riparian Communities on Flooding Events. Riparian communities along smaller headwater streams provide large-woody debris and large-organic debris to stream channels that help reduce downstream floodflows. The streambank stabilization afforded by riparian systems can also reduce the delivery of sediment, thereby maintaining the capacity of the stream channel to transmit floodwaters. However, there are tradeoffs in evaluating the influences of riparian communities on flooding events where dense-woody vegetation such as phreatophytes occurs on a floodplain. Dense stands of phreatophytes can increase the roughness of the floodplain and the streambanks to the point of slowing the velocity of floodwaters and, in doing so, cause higher stages associated with peak stormflows, causing greater flood damages in upstream channels. On the other hand, this increased storage and reduced streamflow velocities on upstream floodplains can reduce flood frequency and damaging flows downstream.

WETLANDS

Wetlands occupy only a small area on most watershed landscapes but their hydrologic role in terms of storage as well as their influence on sediment and water quality are often substantial. Wetlands can be crucial to the hydrologic functioning of the watershed and, therefore, to IWM (Fig. 13.5). Wetlands have other values as well, particularly their unique plant communities that provide valuable wildlife habitat that is the result of prevalent water. While the terms *riparian systems* and *wetlands* are often used interchangeably, the two systems differ from a hydrologic and an ecological standpoint.

Wetlands are areas inundated or saturated by surface or groundwater at a frequency and duration sufficient to support a prevalence of vegetation typically adapted to saturated soil conditions. As a result, wetlands have unique vegetation (hydrophytes) and soils (hydric soils) that distinguish them from adjacent uplands. Riparian areas can be considered as the interface between terrestrial systems and either bodies of water or wetlands. However, wetlands can occur in riparian areas.

The hydrologic functions of wetlands include:

- intercepting and reducing the transport of sediment and other pollutants to downstream water bodies;
- attenuating stormflow peaks;
- supporting hydric plants not found elsewhere on a watershed, thereby providing enriched biodiversity; and
- providing habitats for many organisms from microbes to migrating waterfowl – which has led to wetland preservation programs by sportsmen groups throughout North America.

Wetlands were once considered by many people to be "swamplands" of little economic value unless they are drained to create agricultural croplands or reduce the occurrence of mosquito-borne diseases such as malaria. However, because of their hydrologic, water

FIGURE 13.5. Wetlands are characterized by the permanent or frequent presence of water and occur in a variety of inland and coastal land forms. Pictured is a black ash wetland in northern Minnesota (Photograph by Mark Davidson) (For a color version of this photo, see the color plate section)

quality, habitat, and biodiversity values, efforts to manage, protect, and restore wetlands have become paramount today. The challenge confronting watershed managers is balancing the needed protection or restoration (when required) to retain the above values of these ecosystems while under pressures of agricultural, urban, and other types of development.

Wetland Hydrology

Wetlands are often touted as being important groundwater-recharge zones and systems that attenuate flood peaks and sustain streamflow during the dry season. Likewise, wetlands have been suggested to function like the “kidneys” of a watershed – cleansing or purifying water that enters and then releasing water of a higher quality. As a result of such claims, concerns about the loss of wetlands have resulted in programs to restore or create wetlands. To what extent are such claims true? To address these questions, we will first consider the hydrologic conditions under which wetlands form and then review the hydrologic processes and functions of wetlands under different settings. Finally, we will consider the effects of wetland loss on the quantity and quality of groundwater and flow regimes of streams and rivers.

Hydrologic Conditions for Wetland Establishment

Two hydrologic conditions are necessary for wetlands to form: (1) there is a persistent excess of water at the earth's surface, and (2) the topography and climatic regime result in slow-moving water (Verry, 1997). Excess water occurs at the earth's surface under conditions where annual precipitation exceeds annual *PET*, where groundwater intersects the surface or where there are depressions in the landscape that collect runoff or subsurface flow. Low topographic relief or depressions result in stagnant or slow movement of water, which are conducive to anaerobic conditions and associated microbial communities in soils, resulting in the formation of hydric soils and the development of unique plant communities that are adapted to such conditions. Wetlands typically have a poorly defined or nonexistent channel-drainage system; drainage density is very low in contrast to upland areas of a watershed. As a result of slow-moving water and saturated soils, dead organic material tends to accumulate with the decomposition of organic matter resulting in even further reduced oxygen levels in the zone of saturation.

The type of wetlands that develops is further determined by climatic regime and the source of water that forms the wetland. The source of water, its quality, and periodicity all affect wetland development. Although precipitation is the ultimate source of water to all inland wetlands, the residence time of precipitation in the soil and geologic strata prior to entering a wetland site will dictate water chemistry and the types of plant communities that develop. Regional groundwater-fed wetlands tend to be mineral rich, their development being a function of residence time and the types of geologic strata from which the water originates. Some sources of groundwater are more calcium rich than others, a characteristic that can influence the type of wetland plants that develop. If the water sources are primarily precipitation and surface runoff, the wetland water will generally have lower nutrient content and a lower pH (higher concentrations of H^+) than those that are fed by the more mineral-rich regional groundwater. Depressional wetlands and those with perched water tables are examples of nutrient-poor wetlands. However, surface runoff from agricultural or urban lands can import excessive nutrients that can affect the productivity and species composition of vegetative communities. Furthermore, wetlands occurring near oceans can receive high concentrations of sodium chloride. All these factors affect the types of plants and overall productivity of biomass in wetlands. The vegetation that develops on wetlands can be forests, shrubs, mosses, grasses, sedges, and other hydrophytes.

From the above discussion, it is clear that the role of groundwater is a key factor in determining the type of wetland that develops. Wetlands can develop in dryland regions, where water collects in depressions and maintains a persistent, saturated soil condition. Wetlands most often are low-lying areas that are connected with groundwater. In some cases, wetlands are maintained by their own water table, which in essence is a perched groundwater system. Perched refers to limited downward water percolation and, therefore, limited connections to regional groundwater. In any case, the water table of wetlands is

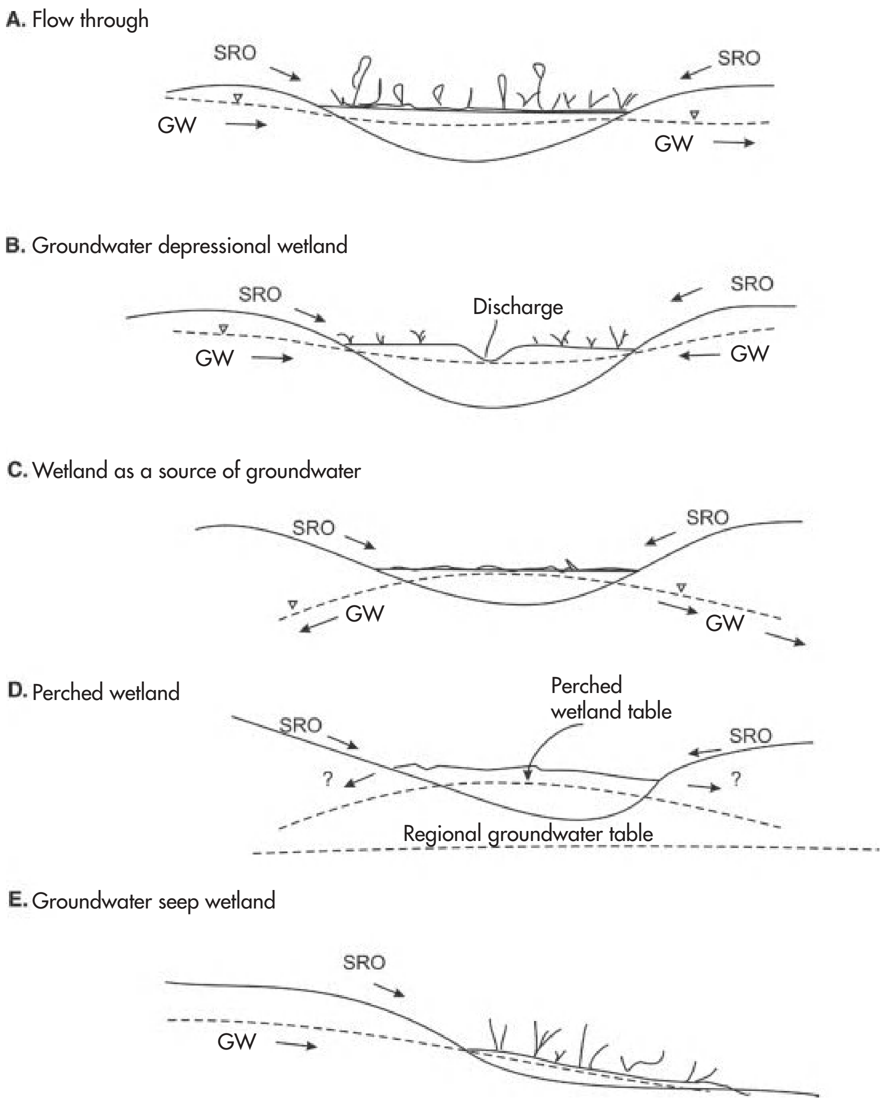


FIGURE 13.6. Surface and groundwater relationships that form wetlands (modified from Mitsch and Gosselink, 1993)

normally above, at, or near the ground surface throughout the year. Examples of the various surface–groundwater relationships for wetland formation are illustrated in Figure 13.6.

The flow-through type wetland (Fig. 13.6A) maintains a fairly stable water table in the wetland throughout the year because of the direct connection with the regional groundwater source. In peatlands, this type would be typical minerotrophic (mineral-rich) fens, which maintain a more diverse plant community than ombrotrophic (mineral-poor) bogs where water tables are maintained by precipitation and perched above the regional water table (as in the case of Fig. 13.6D). Raised bogs can also form on top of fen peatlands, a situation in which a perched water table lies immediately above the regional water table. The vegetation in these raised bogs, however, is not in direct contact with the regional groundwater and as a result exhibits the same mineral-poor characteristics of a perched bog.

Examples of depressional wetland formation are situations where surface runoff and groundwater flow into the depression, and when the water table in the wetland is sufficiently raised, the wetland discharges water from a channel outlet (Fig. 13.6B). In drier environments, a depressional wetland can form from seasonal precipitation excess and surface runoff and either discharge to groundwater (from the edges of the wetland) during periods of high water tables, as in Figure 13.6C, or can maintain a perched water table above the regional groundwater (Fig. 13.6D). A point to consider with depressional and perched wetlands is the linkage between the wetland water table and the regional groundwater table. These wetlands form over time because the downward flow of water is impeded by impermeable strata like dense sediment that was compacted under the weight of glaciers. Therefore, one would not expect to find significant exchanges of water vertically through the bottom of these wetlands. Most likely the greatest exchange between perched water and regional groundwater would occur along the edges of these wetlands during wet periods when water tables are raised (see Chapter 5, Fig. 5.9). If there was an active exchange of water, then one would question how the wetlands were able to form in the first place. For example, depressional wetlands that are elevated above the regional groundwater table would not be able to form under deep sandy soils. Such depressions would represent focused recharge areas for regional groundwater, but by virtue of the fact that there is an active downward percolation of water, the necessary water table and saturated soil conditions necessary to form a wetland would not likely be maintained. Once such depressions fill in with finer sediments, an impeding layer can form over time. A water table could then be maintained, but then the active recharge to regional groundwater would also diminish over time, except along the wetland margins.

Wetlands can also form in areas where springs or seeps occur (Fig. 13.6E). This same process can occur along lakes and coastal areas where the groundwater discharges to the water body through an adjacent, low-lying area.

Water Budget of Wetlands

The linkages and relationships between a wetland and groundwater, as discussed in the previous section, largely determine the water budget and the hydrologic behavior of wetlands. Water-budget components of perched peatland bogs and groundwater-fed peatland fens differ in their accounting for groundwater input and output as illustrated in Figure 13.7. Although these examples are of peatlands, the same water-budget components would apply to other perched and groundwater-fed wetlands.

Wetlands that are fed by regional groundwater exhibit a relatively stable water table. If there is a channel outlet in the wetland, these systems will sustain a relatively even pattern of streamflow throughout wet and dry seasons. In contrast, wetlands that have perched water tables and a channel outlet can be separated from regional groundwater, exhibiting greater seasonal fluctuations in both the water table and the streamflow discharge. Examples of the seasonal pattern of streamflow from a perched bog and a groundwater-fed fen are illustrated in Figure 13.8. Flow duration relationships for these two peatland types are illustrated in Figure 13.9.

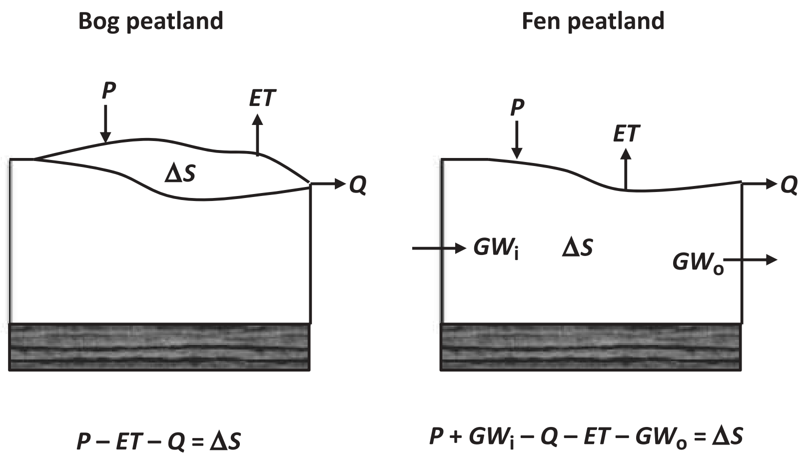


FIGURE 13.7. Water budgets for a perched peatland bog and a groundwater-fed peatland fen. *P*, precipitation; *ET*, evapotranspiration; *Q*, surface discharge; ΔS, change in storage; *GW*, groundwater

Wetland water-budget components of *ET* and outflow, whether as surface discharge to streams or as groundwater discharge, are a function of water-table depth. When the water table is located above, at, or close to the surface, both *ET* and outflow are large. As the water-table elevation falls, the water table becomes deeper and *ET* and discharge from the wetland diminish. The relative magnitude of water-budget components for different wetlands vary, but in general, wetlands exhibit high-annual *ET* losses and a corresponding low-annual water yield as discharge to either surface or groundwater systems. In northern latitude wetlands, snowmelt and rain-on-snow events in the spring are normally the periods of greatest discharge to streamflow. During the summer months, high *ET* rates drop water-table elevations and reduce flow from wetlands except for rainfall events with large volumes and high intensity.

Evapotranspiration. Wetlands evaporate and transpire water at or near the potential (*PET*) rate during the growing season. The readily available water supply for plants and wet-surface conditions that characterize most wetlands explain this high rate of *ET*. When the water table is close to the wetland surface, *ET* can exceed the evaporation from an open-water surface. Although field measurements of *ET* are sparse, examples of measured rates are listed below:

- Daily *ET* rates measured in northern Minnesota peatlands averaged 3–4.5 mm/day with maximum rates of 6.0 mm/day during the growing season based on measurements from evapotranspirometers and the Bowen ratio–energy balance methods (Brooks et al., 2011). *ET* often exceeded pan evaporation rates (Bay, 1966) and equaled Thornthwaite *PET* estimates during the growing season (Verry and Timons, 1982).
- Daily *ET* for 6 days in late summer from a mountain bog in Wyoming averaged 3.3 mm/day in contrast to average daily pan evaporation rates of 2.6 mm/day (Sturges, 1968).

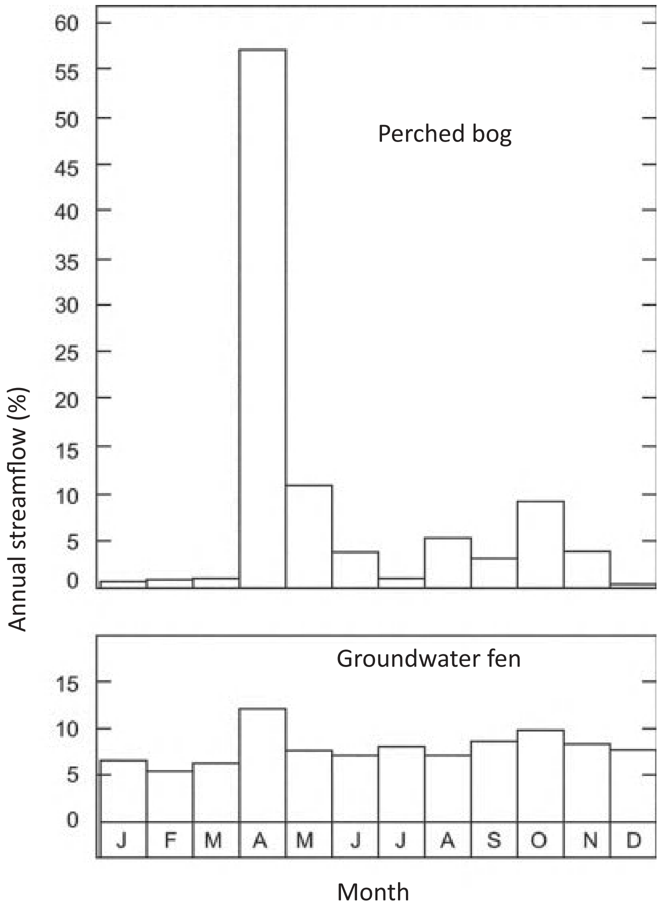


FIGURE 13.8. Seasonal pattern of monthly streamflow from a perched bog and a groundwater-fed fen in northern Minnesota (from Boelter and Verry, 1977)

- Comparisons of a bog in Quebec showed that *ET* averaged 2.9 mm/day and ranged from 1.9 to 3.6 mm/day (Van Seters and Price, 2001).
- *ET* rates from three reed (*Phragmites australis*) sites in England were quite variable, ranging from 1.0 to 6.3 mm/day over a 4-year period (Fermor et al., 2001).

The ratio of *ET*/*PET* for peatlands is close to 1.0 when the water table is between the surface and to a depth of about 25 cm (Verry, 1997). It is somewhat lower when the water table is above the surface and diminishes dramatically when the water table drops below around 30 cm depth. For one of the peat bogs used in this relationship, the water table was found to reside between the surface and 30 cm depth more than 85% of the time, based on a 27-year record (Verry et al., 1988). This tells us that growing-season *ET* of peatlands will occur at the potential rate in all but drought years.

For wetlands such as the prairie-pothole wetlands of North America, water tables can fluctuate dramatically from year to year and seasonally and, therefore, would be expected to have more variable *ET*. Potholes can become completely dry during periods of drought,

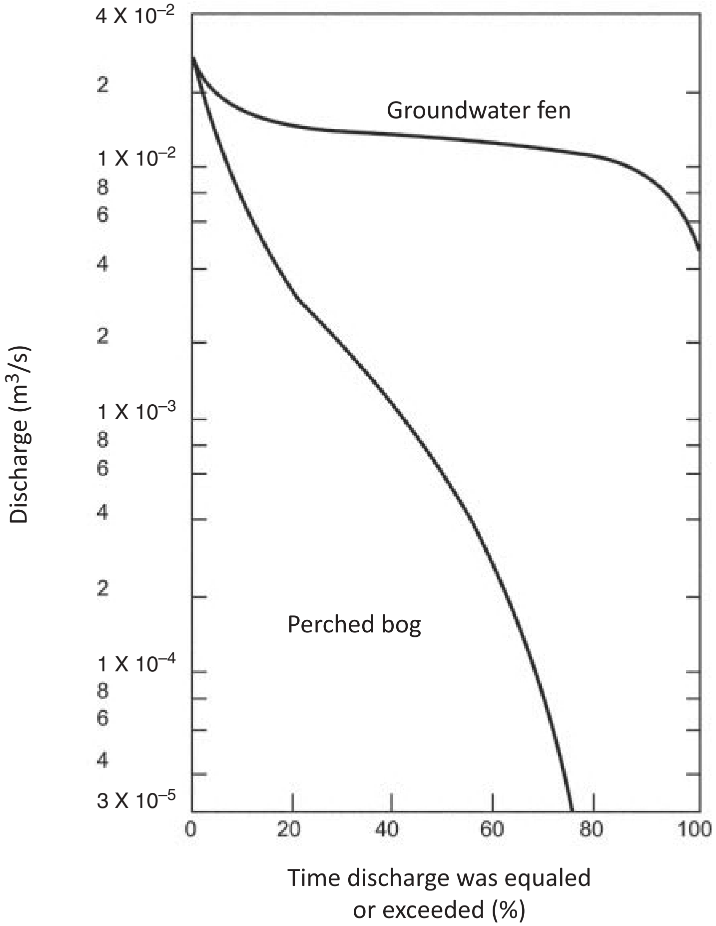


FIGURE 13.9. Streamflow duration relationships for a perched bog and a groundwater-fed fen watershed in northern Minnesota, each of about 53 ha area (from Boelter and Verry, 1977)

greatly reducing *ET*. Under normal precipitation regimes, *ET* can be high and a relatively large component of the water budget. For example, from May through October, pothole wetlands in North Dakota lost 604 and 682 mm from potholes with emergent and limited emergent vegetation, respectively (Shjeflo, 1968). Average precipitation for the period of study was 418 mm/year. In these cases, the emergent vegetation seemed to limit total *ET*.

ET measurements from other types of wetlands indicate variable relationships between vegetated wetlands and open bodies of water. For example, *ET* from a vegetated wetland in New Hampshire was 80% greater than that from an adjacent open body of water (Hall et al., 1972). In contrast, *ET* from cypress wetlands in Florida was lower than evaporation from open water (Brown, 1981).

Groundwater Recharge. Groundwater recharge from most types of wetlands is limited and generally not a large component of their water budgets. In northern prairie wetlands of North America, water is primarily lost by means of *ET* and to a lesser extent by seepage. Isolated wetlands can recharge groundwater by lateral seepage (Kleinberg, 1984). However, seepage is often difficult to quantify and can be largely lost through *ET* from the wet edges of prairie potholes (Su et al., 2000). Recharge to regional groundwater aquifers from these wetlands has been reported to range from 2 to 45 mm/year (van der Kamp and Hayashi, 1998). The higher rates would be expected only where there are large hydraulic gradients from the wetlands to the groundwater aquifer, where hydraulic conductivities are not restrictive, and in locations where the aquifer is shallow.

Based on how they are formed over time, prairie potholes and similar wetlands would be expected to recharge groundwater more from their edges during periods of high water tables than from percolation through bottom deposits. Even in wetlands where the water table is maintained throughout the year, the rates of recharge would be expected to be low. This rationale follows when one considers how and why wetlands form in the first place, as discussed earlier. If they are not groundwater-discharge areas, wetlands are the result of an impeding layer that restricts the downward percolation of water, suggesting that the saturated hydraulic conductivities (K_{sat}) of soils underlying wetlands are minimal.

Organic material in peatlands accumulates over a long period of time (centuries) during which the slow anaerobic decomposition of the lower level peat breaks down the coarser material to finer particles. As the organic matter deposits become thicker, the lower layers become more compressed; consequently, K_{sat} diminishes rapidly with depth into the soil, which corresponds to the increased decomposition of peat. The von Post degree of decomposition (von Post and Granlund, 1926) is used to characterize peat soils. Figure 13.10 illustrates that the higher the value on the von Post scale (ranging from H1 to H10),

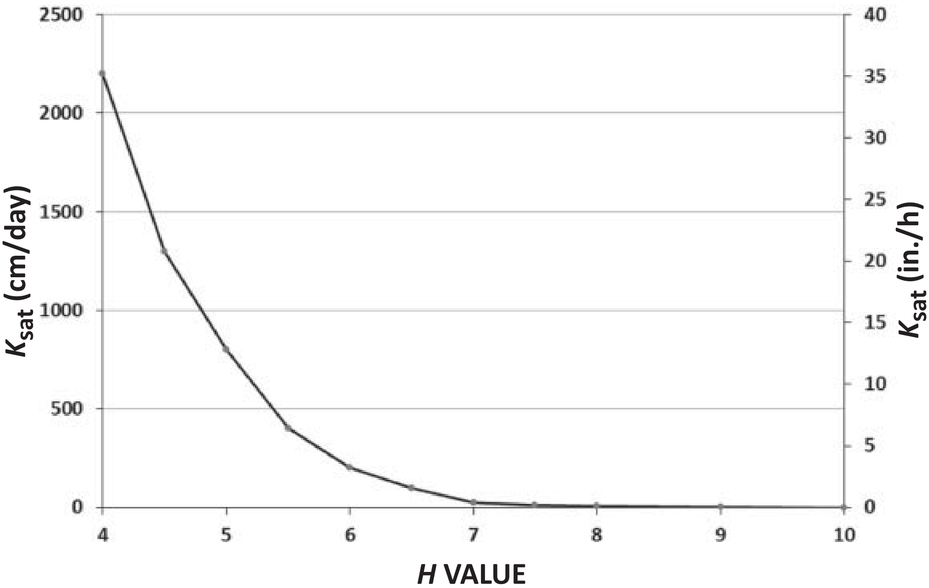


FIGURE 13.10. Relationship between saturated hydraulic conductivity (K_{sat}) and von Post H values of peat decomposition. The relationship for H4–H7 was determined by Gafni and Brooks (1990) and for H8–H10 from Päivänen (1973), as presented by Verry et al. (2011) (copyright by CRC Press, by permission from Taylor and Francis Group LLC)

the greater the decomposition and the lower the K_{sat}. As a result, vertical, downward flow of water through the peatland is restricted with depth. Water is then forced to flow in a horizontal direction in the peatland in response to the hydraulic gradient. Furthermore, the hydraulic gradients remain quite constant throughout the year, even though water levels dropped over the growing season (Gafni and Brooks, 1990).

Flow velocities through peatlands and wetlands in general tend to be very slow because of the low hydraulic gradients, even where K_{sat} is relatively high (refer to Darcy's Law). In contrast, if a greater hydraulic gradient in the soil profile is imposed, velocities in the deeper horizons would remain slow because of the lower K_{sat} in the deeper layers. It is not uncommon to see expressions of vertical pressure gradients within wetlands based on piezometer measurements that suggest vertical water movement into and out of wetlands. Interpreting such relationships in determining rates of water flow again requires that K_{sat} of the deeper layers of wetland soils be carefully evaluated.

Runoff and Streamflow from Wetlands. The depth of the water table governs wetland runoff responses to rainfall and snowmelt. Wetlands with a channel outlet normally yield high amounts of runoff only when the water table is at or above the soil surface. The percentage of rainfall that becomes streamflow is usually small except during the rainy season or snowmelt season in temperate climates and when plants are relatively dormant. High *ET* rates during the summer lower the water table, thereby creating storage in the wetland that must be satisfied before water tables rise and discharge can increase. In the case of perched wetlands with outlets, it is not uncommon for streamflow to cease late in the growing season or during droughts. An example of the cumulative frequency in which

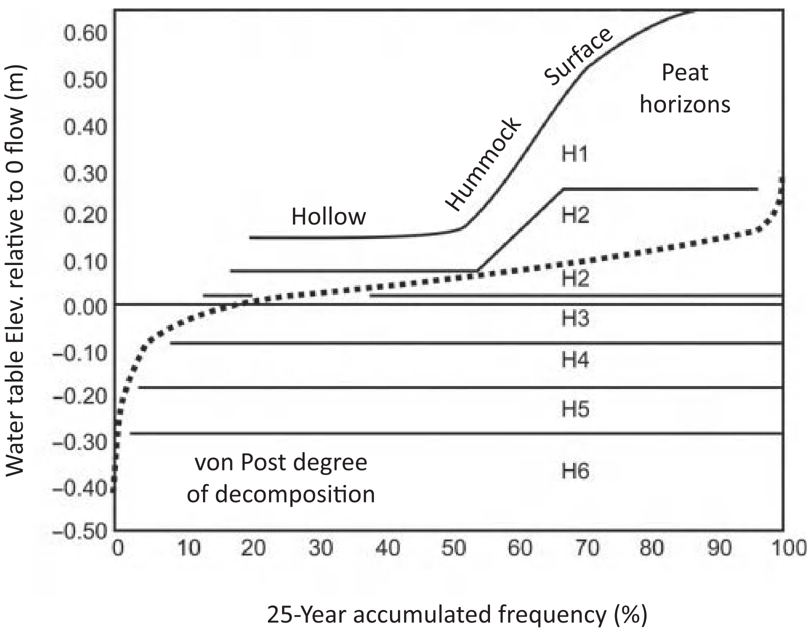


FIGURE 13.11. Cumulative frequency of occurrence of the water-table elevation at different peat horizons in a perched bog in northern Minnesota (Verry et al., 1988)

the water table occurs at different depths in a peatland bog is presented in Figure 13.11. The corresponding discharges to these water-table elevations indicate that 25% of the time, there is no discharge (corresponding to depths of 14.5 cm or more below the hollow), and significant discharges from the bog occur only 5% of the time when the water table exceeds the hollow elevation. About 70% of the time, the bog discharges water to the stream channel, but at very slow rates.

Depressional wetlands, peatlands, and the like provide storage benefits and hydrologic functions that are similar to a shallow reservoir. Largely because of their flat topography and lack of well-defined channels, most wetlands attenuate flood peaks by temporarily storing or detaining water (Fig. 13.12). There must be an established relationship between storage and water-table elevation and water-table elevation and discharge for this analogy to hold. The difference between the two systems is that (1) the wetland relationships are based on water-table elevations in contrast to reservoir pool elevations, and (2) storage

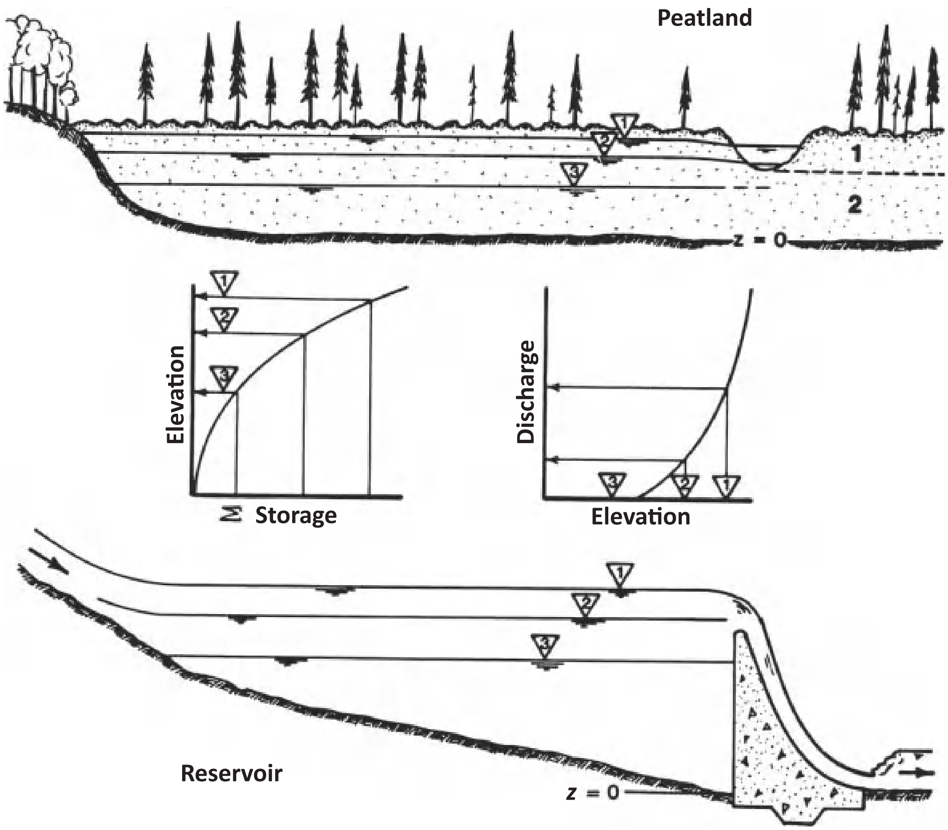


FIGURE 13.12. Streamflow from a peatland is governed by water-table elevation similar to the way that discharge from a reservoir is governed by pool elevation (Guertin et al., 1987). Point 1 = high storage and elevation (peatland) correspond to high discharges (reservoir); point 2 = as water table (peatland) or pool (reservoir) drops, discharge decreases; point 3 = when respective elevations drop below the outlet (peatland) or spillway (reservoir), discharge ceases; z = datum

in wetlands is a function of wetland–soil characteristics whereas storage in reservoirs is free, unbound water. Storage–water-table elevation relationships can be determined for any wetland. Developing a water-table elevation–discharge relationship for wetlands with an outlet channel is more problematic, but has been accomplished as described in Box 13.3. The development of such relationships facilitates modeling of streamflow from wetlands which have outlet channels. Relationships between the water-table elevation, as measured by a well, and outflow from the wetland need to be developed from field data.

The suggested analogy between wetlands and reservoirs suggests that flood storage–flood-control benefits can be attributed to wetlands. Although storage is more limited than with lakes or reservoirs, the presence of wetlands can influence peak-flow discharge from watersheds. Up to a point, the greater the percentage of watershed area composed of any combination of wetlands, reservoirs, or lakes, the greater the reduction in peak-flow discharges for a given storm event. In northern Wisconsin such benefits have been considered significant when up to 20–30% of a watershed has a combination of wetlands, lakes, or reservoirs as illustrated in Figure 12.13 (Conger, 1971).

Concluding Thoughts – Hydrologic Functions of Wetlands

When considering the hydrologic consequences of human interventions on wetlands, or conversely, how wetland restoration can affect watershed hydrology, it is useful to consider the following characteristics and response of natural, undisturbed wetlands:

- Shallow water tables, flat topography, and low drainage densities are dominant features of most wetlands.
- The depth of the water table governs *ET* and streamflow discharge from wetlands.
- Annual *ET* far exceeds annual discharge for most wetlands.
- Wetlands tend to be groundwater-discharge areas more often than groundwater-recharge areas.
- Largely because of their flat topography, wetlands function much like simple reservoirs that attenuate flood peaks by temporarily storing or detaining water.
- Wetlands linked to regional groundwater systems exhibit less seasonal fluctuation in water table and streamflow discharge than do wetlands that are perched or otherwise isolated from regional groundwater.

When wetlands are altered or converted to some other land use such as urban areas or agricultural cropland, the volume, pattern, and pathways of water flow can be significantly affected as discussed with examples later in this chapter. In addition, there can be significant changes in water quality.

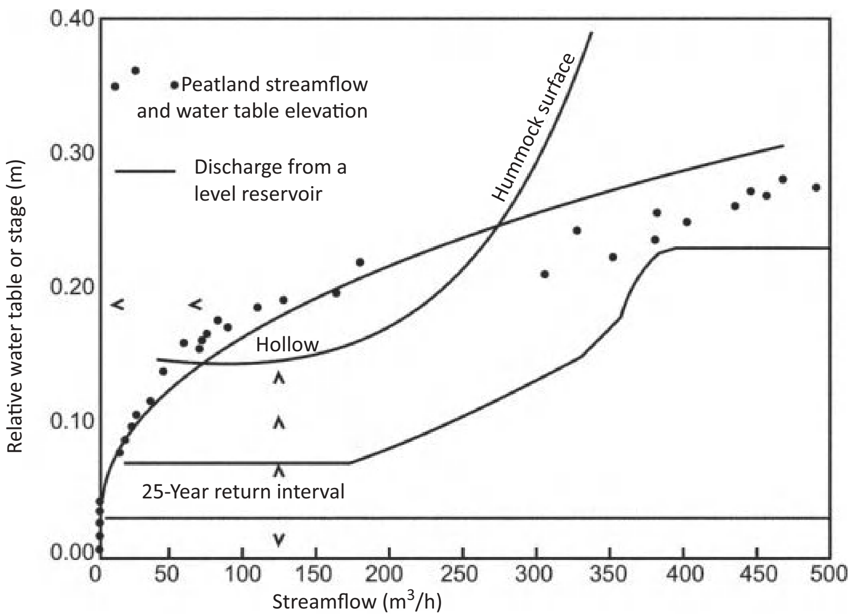
Wetlands and Water Quality

A comprehensive review of wetlands and water-quality relationships is beyond the scope of this book. However, a general overview of how wetlands influence certain water-quality characteristics is presented below.

The physical location of wetlands in most watersheds suggests that they are largely receivers of water from precipitation, surface runoff, and/or groundwater. The water chemistry of a wetland reflects the source of water. For example, fen peatlands in northern

Box 13.3

Streamflow Response from a Peatland Similar to that of an Unregulated Reservoir (from Verry et al., 1988)
Water-table elevation–discharge relationships for a 3.24 ha peat bog in northern Minnesota was similar to those predicted for a small, unregulated reservoir (refer to Fig. 13.12). When the water-table elevations in the bog were modified to reflect the specific yield from the peat soils, they matched closely with the same discharges that would occur from a level reservoir (see the figure below). Departures from the reservoir elevation–discharge relationship occurred when the water table in the bog rose slightly above the elevation of the hollows of the hummock-like peat surface (the dots in the figure were greater than the solid line of the reservoir relationship). The higher elevation–discharge relationship in this 75–100 m^3/h range is attributed to the resistance to flow offered by the hummocks and trees in the bog. As the water-table elevation is raised further, the discharge rates from the bog were higher than would occur from a level reservoir. These higher flow rates were attributed to channel-like flow in the lagg of the peatland in which a moving wedge of water has a greater velocity than that of a flat reservoir surface.



Measured relationships between water table elevation and discharge from a bog and similar relationships for a shallow reservoir with the same dimensions (from Verry et al., 1988)

The above reservoir relationships determined from field measurements were incorporated into a hydrologic model capable of predicting streamflow discharge from peatlands (Guertin et al., 1987). The Peatland Hydrologic Impact Model (PHIM) developed at the University of Minnesota computes water-budget components so that storage conditions in the peatland are determined. Storage–elevation and elevation–discharge relationships are then incorporated into the model to predict streamflow discharge from peatlands.

Minnesota are expressions of regional groundwater and have a higher mineral content and higher pH than rainfall-fed bogs (Clausen and Brooks, 1983). The higher mineral content causes the specific conductance of water discharging from fen peatlands to be significantly higher than bogs and, as a result, can be used to identify both the source of water and the peatland type.

Wetlands that receive overland flow or stream discharge from upland watershed areas are often considered to be sinks in terms of water-quality constituents. The velocity of any flow of water into wetlands is reduced, resulting in sediment deposition. Along with this sediment deposition are minerals, phosphorus, and sometimes heavy metals. Many of the dissolved nutrients that enter a wetland can be taken up by wetland plants, resulting in lower nutrient loads leaving a wetland than that which entered the wetland. It is this “cleansing” function that is often seen as a valuable benefit of wetlands on a watershed. Whether this is true depends on the water-quality constituent and the type of wetland. Generalities concerning wetland effects on water quality include the following:

- Sediment and particulate matter deposited in wetlands can reduce that which would otherwise enter a stream system. The capacity of a wetland for storing particulates

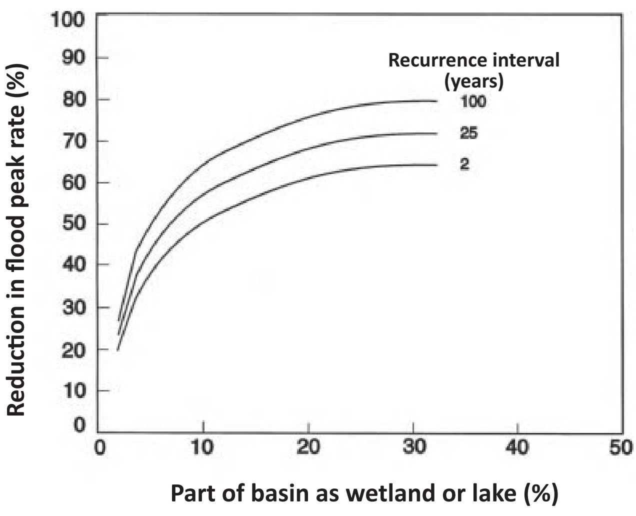


FIGURE 12.13. Relationship between peak-flow discharge (%) and percentage of a watershed in wetlands or lakes for different recurrence intervals in northern Wisconsin (adapted from Conger, 1971, as presented by Verry, 1988)

is finite. However, in the case of particulate phosphorus (P) that is deposited over time, some P can eventually be converted into soluble forms which can then leave the wetland and enter streams or groundwater systems in a soluble form.
- Nitrogen entering a wetland can be taken up by plants and, because of anaerobic conditions, processes of denitrification in wetlands can convert NO_3 to gaseous nitrogen (N_2). These processes can reduce nitrogen available for export to streams and groundwater systems.
- The role of wetlands in removing excess nutrients from upland runoff is a function of the types of plants present, the total biomass, and the fate of these plants. Eventually, plants will die and the nutrients taken up will be recycled. Therefore, some type of management that removes and regenerates vegetation or a natural process such as fire (see below) is needed to sustain a nutrient "sink" function.

Because of such functions, wetlands are being constructed or rehabilitated in riparian areas to intercept and remove certain constituents before they enter a stream or lake. Magner and Alexander (2008) describe how interceptor wetlands can be used to treat impaired water quality in agricultural landscapes (examples will be further discussed in Chapter 14).

Artificial wetlands are being created in urban areas to provide these functions and are designed to treat urban runoff before it enters streams or lakes. Likewise, restored and created wetlands in agricultural areas have been used to reduce nutrient export to streams.

Wetlands and Integrated Watershed Management

Wetlands are integral components of many watersheds and play valuable roles influencing groundwater–surface water interactions, surface water flows, and water quality (discussed in Chapter 7). Management actions that alter or remove wetlands must be understood in terms of their consequences to groundwater and to the flow and quality of surface water. Furthermore, the benefits of rehabilitating, restoring, or creating wetlands on a landscape need to be understood in terms of their influence on water and their role in IWM.

Management of wetlands is often "custodial in nature" to achieve their needed protection while retaining their hydrologic values. In wildland watersheds, active management of wetlands includes

- cutting trees for commercial purposes or influencing hydrologic functioning;
- controlling wildfires to minimize their impacts;
- applying prescribed fire to achieve management goals; and
- mitigating the effects of roads.

From a watershed hydrology perspective, the types of changes to wetlands that would cause the greatest hydrologic change include

- draining of wetlands for economic development;
- mining peatlands for horticultural products or energy purposes;
- removal of wetlands; and
- restoration or creation of wetlands.

These types of changes are discussed to provide a hydrologic perspective for planning and managing wetland ecosystems and watersheds to achieve specific objectives of IWM or avoid excessive flows of water, prevent impaired water quality, and avoid detrimental effects on groundwater–surface water interactions.

Cutting Trees

Trees in forested wetlands are sometimes cut for commercial purposes. Special considerations for harvesting operations and management are needed to maintain the integrity of the wetland to the extent possible. Protecting soils and water quality is of particular concern in forested wetlands. If possible, tree-cutting activities should be scheduled in the dry season or when the soils are frozen as in the northern latitudes to limit their impacts on the soil.

Winter harvesting of wetland forests, such as black spruce (*Picea mariana*), is common in the northern United States and Canada. Clearcutting of black spruce is more common than cutting in strips or patches, and most peatland conifer stands need 20–25 years to reach full stocking (Grigal and Brooks, 1997). Winter harvesting by stripcutting and clearcutting of black spruce in a peat bog and clearcutting in a groundwater-fed fen did not change annual streamflows in north-central Minnesota (Sebestyen et al., 2011). The hydrologic conditions of these peatlands were resilient to tree harvesting when surrounding upland forests were not harvested. However, the streamflow pattern from peat bogs is more erratic than flows from fens due to the dependence on precipitation and the absence of regional groundwater flows.

Follow-up watershed management after trees are harvested often includes attempts to artificially regenerate the wetland with trees when natural regeneration is unlikely. This being the case will often require some form of site preparation before planting. The disturbance to the surface soils should be minimized regardless of the site-preparation technique used.

Controlling Wildfires

That wetlands can burn almost seems contradictory. However, a high-severity wildfire can consume much if not all of the vegetation in a wetland ecosystem while also impacting the hydrologic properties of the soil. Wildfires in peat bogs in a drought period are especially devastating. A smoldering ground wildfire that continues to burn in the organic soils for a long time under drought conditions will modify the water-holding capacity of peat soils. Such fires can continue until the water table is restored or the drought conditions end. When the wildfire also burns on the surrounding watershed, increased surface runoff can be generated that in turn increases water flows and sediment movement into embedded wetland ecosystems.

Suppression activities undertaken to control a wildfire can in themselves affect the hydrologic functioning of a wetland ecosystem especially when heavy equipment such as tracked bulldozers are used to construct firelines. A watershed manager in consultation with a fire-control specialist should insure that the impacts of suppressing the fire are minimal.

Prescribed Burning

Prescribed burning can be a valuable tool in the management of wetlands. The benefits include

- removing accumulations of dead herbaceous plants to prevent a buildup of debris;
- reducing or eliminating undesirable woody plants that have invaded into the wetland;
- bringing about a preferred pattern of plant succession;

- cleaning basin impoundments before possible flooding; and
- creating nesting areas for waterfowl (Kirby et al., 1988).

Most of the prescribed burning in wetlands takes place in the dry season when the moisture content of flammable fuels is low enough for a successful ignition of the fire.

Naturally occurring lightning-caused fire ignitions of low intensities have helped to maintain the species composition, structure, and general heath of wetland ecosystems in some instances.

Mitigating the Effects of Roads

Roads, road-related drainage systems, and stream-channel crossings can individually or collectively impact the hydrologic functioning of a wetland ecosystem by intercepting and concentrating or diverting surface runoff; intercepting and diverting subsurface flows of water; and initiating and accelerating soil erosion. Changes in surface–groundwater flow pathways and exchanges can result in loss of wetland vegetation and function. Watershed managers should also be concerned about the effects of road-generated sediment and turbidity on downstream water bodies.

Road construction and maintenance practices can be modified to favor the rehabilitation of sites previously damaged by human-induced disturbances such as livestock grazing or tree harvesting. It is fortunate that BMPs have been identified to mitigate the magnitudes of sediment and other pollutants originating from road construction and maintenance (see Chapter 14). Site-damaged road segments can be relocated, modifications or replacement of stream-crossing structures by culverts, fords, or bridges are possible, and modifications of ditch systems and cross-drains can disperse captured flows (LaFayette et al., 1992).

Wetland Drainage for Economic Development

Wetlands have been and continue to be drained, altered, or eliminated for a variety of purposes such as increasing agricultural production, urban development, and other forms of economic development. Much of this development requires drainage to remove the excess water in the wetland ecosystem. However, the hydrologic implications of removing this excess water have not always been clear, although reducing or increasing the percentage of a basin in wetlands or lakes has been shown to affect streamflow peaks as shown earlier in Figure 12.13. The physical and biological effects of draining wetlands are often assessed in terms of the changes in storage and routing of water within the surrounding watershed to better understand the impacts on the pattern of streamflows and possible flooding emanating from the drained sites. Changes in water quality can also be related to these changes in streamflow regimes.

Although much of the emphasis in wetland drainage is associated with removing excess water, there are examples of altering wetlands to enhance downstream flows. An example of altering a major wetland for this purpose is the Jonglei Canal proposal to divert Nile River flow from a portion of the large Sudd wetland in Sudan to increase flows in the arid lower Nile Basin (Box 13.4). Implementation of the project has been halted because of the scale of the project and adverse environmental and social impacts in the southern Sudan region.

Box 13.4

Proposals to Divert Water from the Sudd Wetland to Enhance Water Supplies in the Lower Nile River

Schemes to divert parts of the Nile River to increase water supplies in the Lower Nile date back to ancient times (see Box 1.1). In 1946 the Jonglei Canal was proposed by the government of Egypt to divert part of the flow in the Nile River. The proposed diversion would have routed water around the 30,000 km^2 Sudd Wetland in southern Sudan for the purposes of increasing flow to the Lower Nile Basin (Howell et al., 2009). Increased flows of about 7% would be expected in the Nile River flow to Egypt with the proposed diversion. The reason for this increased flow is that 55% of all water entering the Sudd wetland is lost via *ET* (Baecher et al. 2000); much of these losses would be avoided with the diversion. The diversion has only been partially completed although proposals to restart the project have been discussed as recent as 2008 (see Wikipedia website, accessed January 5, 2011). However, the completion of the diversion could have undesirable impacts (Baecher et al., 2000), including:

- The proposed diversion would reduce the size of the wetland by around 35%. The Sudd Wetland plays an important role in storing floodwaters and trapping sediment from the White Nile River which ultimately benefits communities in the Lower Nile River and Lake Nasser.
- The seasonally flooded grasslands of the Sudd provide critical habitat for wildlife, a rich biodiversity, and valuable land used for grazing and rain-fed agriculture that supports the rural poor inhabitants of the region.

A question to be addressed with the Jonglei Canal proposal is whether the benefits of added flows downstream outweigh the costs in terms of the loss of production and environmental services attributed to the Sudd Wetland. Taking an IWM approach in making assessments of the costs and benefits of the proposed project (described in Chapter 15) would be warranted before continuing the project.

Wetland Drainage for Agriculture

Widespread drainage of wetlands has occurred in the United States and other parts of the world to increase the land area for agricultural production. While there is little question that food production has increased as a result of draining wetlands for agricultural purposes in many areas, what is less clear are the hydrologic, water quality, and other environmental implications of this drainage. There is concern that increased agricultural production at the expense of the loss of wetlands and the resulting hydrologic functions and impaired water quality in many areas is not sustainable. Before we can develop IWM practices to address these issues, the cause-and-effect relationships of wetland drainage and loss must be understood.

One of the more widely known wetlands in the United States that has undergone drainage and development is the Everglades of southern Florida. The scale of drainage and the following development of land for agriculture and other purposes in the Everglades have led to significant hydrologic and ecosystem changes that have become controversial (Box 13.5). As with the concerns with the Sudd wetland (discussed in Box 13.4), the drainage of one of the largest wetlands in the United States has become a major battlefield between developers and environmentalists. Solutions to such issues are not often of a technical nature but require accompanying changes in policies (discussed further in Chapter 14).

In contrast to the scale of the alterations of the Everglades, there has been more of a gradual change in wetlands across the Midwestern United States following European settlement and the expansion of agriculture. Rather than one large wetland being altered, there has been a cumulative loss of many small wetlands extending from the states of Ohio and Illinois to the prairie-pothole region of the Upper Midwestern United States and Canada. In the past half century there has been an expansion of artificial drainage and loss of wetlands throughout the region that has altered the landscape and the hydrologic functions of watersheds.

Drainage in the Minnesota River Basin (MRB) has eliminated more than 90% of the original prairie wetlands (Leach and Magner, 1992) and has changed the hydrologic regime in several ways. Much of the water previously stored in the wetlands that was mostly lost to the atmosphere through *ET* now flows from a watershed through the subsurface drains and discharges directly into stream channels or other water bodies (Fig. 12.14). Lesser amounts of seepage would be converted into quicker water flow from the wetland site. The combination of the loss of storage and the improved conveyance systems for water to leave the watershed would explain the findings in Iowa where similar drainage and vegetative cover changes have increased streamflow and baseflow from rainfall and snowmelt runoff (Zhang and Schilling, 2006; Schilling et al., 2008). Because many of the original prairie-pothole wetlands had no surface outlet, the landscape had many small internally drained and isolated basins. The process of drainage connected many of these basins directly to surface-stream channels that in effect expanded the contributing area of the watershed for existing streams, which in turn drove channel enlargement. Modeling studies (Miller, 1999; Ennaanay, 2006) indicate that the cumulative effects of agricultural development, wetland loss, and the expanded watershed areas due to drainage have also increased stormflow peaks associated with recurrence intervals of 1.5–20 years.

The impact of draining wetlands on water quality can also be significant. Installation of outlet tiles results in bypassing the riparian communities of a watershed in many instances with the tiles routing surface runoff directly into downstream ditches and stream channels. As a consequence, there is little opportunity for nutrient uptake by the riparian vegetation. The loss of wetland functions in terms of denitrification and storage of sediment also has important ramifications downstream. Drainage-induced increases in baseflow and stormflow peaks can also increase the amount of energy available for streambank and channel erosion. If the frequency and duration of bankfull (or near bankfull) flow increases over time, there is greater energy available to work on stream banks and channel sediments. The increased energy of flows can then result in increased stream channel instability, accelerated

Box 13.5

Development Impacts on the Everglades of Florida, USA (from Enzler, 2012)

The Everglades ecosystem is a unique grass-covered wetland extending more than 165 km^2 miles in length and 80–125 km^2 in width from Lake Okeechobee to the Gulf of Mexico and was largely undeveloped until the middle of the nineteenth century (Stevens et al., 1994). There is abundant biodiversity even though the vegetation is largely grasses. The wetland is nutrient-poor as it derives most of its nutrients from rainfall and consequently has low levels of phosphorus and other nutrients.

By 1920 four major canals drained parts of the Everglades for agricultural development, including sugar cane. Urban expansion into the Everglades was facilitated in 1928 with the construction of flood-control dikes that also facilitated the conversion of large areas of the northern Everglades to sugarcane. To protect remnants of the Everglades, the US Congress authorized establishment of the Everglades National Park in 1935. However, the major flood of 1947 and damages to the expanded areas of sugarcane production triggered the massive Central and South Florida Flood Control (C&SF) Project that followed the passage of the Flood Control Act of 1948 by Congress (Light et al., 1995). As a result, much of the Everglades was compartmentalized by levees accompanied by pumping stations and channelization of the Kissimmee River, and control outlets essentially subjected practically all flow in the Everglades to human control. Water from Lake Okeechobee no longer flowed into the southern Everglades but was diverted directly to the Gulf of Mexico or the Atlantic Ocean. As a consequence, urban and agriculture development expanded to occupy more than one-half of the historic Everglades.

The hydrologic consequences of the C&SF Project included:

- changes in the volume and timing of water flowing through the wetlands;
- changes in the pattern of flow from sheet flow to pulsed flow;
- unnatural pooling of water in some locations and over-drainage in others; and
- increased phosphorus concentrations.

The resulting impacts on the National Park included:

- loss of feeding habitat for wading birds;
- abandoned nesting areas in the park;
- extinction of seven physiographic landscapes indigenous to the Everglades; and
- depletion of organic soil accumulations due to rapid oxidation.

More than 30 years after the C&SF project began Congress passed the 2000 Water Resources Development Act that authorized $7.8 billion to restore the Everglades in an effort to recover the lost biodiversity and intrinsic values of the Everglades damaged by human-induced hydrologic changes.

FIGURE 12.14. The drainage of wetlands has removed excess water from a large portion of the prairie wetlands in the Minnesota River Basin for agricultural purposes (photograph by Mary Presnail)

streambank erosion, and increased suspended sediment loads and higher turbidity (Magner et al., 2004). The cumulative effects of agricultural drainage and other changes in land use and stream-channel modifications are discussed in Chapter 14.

Reversing the hydrologic and water-quality impairments caused by conversions of wetland to agricultural croplands require an IWM approach that addresses land-use changes across a watershed including the type of vegetative cover, wetland and riparian system restoration, and the use of BMPs.

Peat Mining

Many peatlands in North America and northern Europe have been cleared with the peat soil extracted for horticultural products or energy purposes. The hydrologic effects of this mining activity include a lowering of the water table and increase in the flow of water from mined areas. The effects of peat mining can also impact the hydrologic functioning of the affected wetland ecosystem.

The mining of a peatland involves a sequence of clearing the overlying vegetation, draining the peatland, and finally extracting the peat soil. Clearing the overlying vegetation and the consequent lowering of the water table reduces *ET* losses and, therefore, increases the amount of water flowing from the mined area. As a result of these processes, surface runoff from the harvested peatland can be increased up to nearly 25% in comparison to the often negligible runoff from a natural bog (Van Seters and Price, 2001).

The ditching process and its connection to a stream outlet will release water that would otherwise be stored in the peat or flow slowly in response to the natural hydraulic gradients. Small storage ponds can be installed to mitigate these surges in flow. Ditching and the excavation of the peat steepen the hydraulic gradients and, in doing so, alter the direction of groundwater flow adjacent to the mined site. Installation of a ditch around the peatland to be mined can change the direction of groundwater flow and, in effect, increase the area contributing to the flow of water at the outlet of the wetland area. Drainage ditches often convey the water more quickly to downstream areas, picking up and transporting organic sediment downstream. Surface-peat soils can also become hydrophobic in the summer, which can also increase surface runoff and sediment delivery from the peatland. Peat soils in mined areas freeze as concrete frost in the winter months, resulting in minimal infiltration for snowmelt and early spring rains. The net effects of these changes are increased stormflow volumes and peak flows and higher levels of sediment from mined areas. Detention ponds are needed to mitigate storm surges and to provide settling basins for organic sediments.

CUMULATIVE EFFECTS

Losses or partial losses of riparian communities and wetland ecosystems and their restoration efforts need to be assessed within the context of an entire watershed landscape and through time to appreciate the cumulative watershed effects (CWE) of these losses. In addition to the negative downstream and other "offsite" effects of riparian and wetland degradation, we also need to recognize the positive effects of rehabilitation or restoration. There are intrinsic values that are lost when riparian and wetland systems become degraded or destroyed that include valuable habitat for wildlife, unique plants and ecosystems, and biodiversity. Such losses are often the result of CWE and can provide the justification for restoration in and of themselves.

The external effects of altered riparian and wetland systems in a watershed are most often viewed in hydrologic terms. It is likely that the CWE of the losses of riparian areas and wetlands in the upper Mississippi River Basin have contributed to higher baseflow and to the magnitude of floods up to recurrence intervals of at least 20–30 years. In many instances these increased flows can be compounded by increases of flows resulting from upland watershed changes such as timber harvesting, urbanization, and conversion from perennial vegetation to annual crops. These increased flows can add to the loading of nonpoint source pollutants to stream channels and can alter flow regimes that impact fish and wildlife. In addition to the expansion of agricultural croplands and changes in farming practices in former prairie and grassland systems, there has also been a conversion of upland and riparian forest cover to croplands, pastures, and urban areas, all of which can have compounding effects on water flow and water quality. Watershed managers need a wide-angle lens to gain the perspective needed to address all elements contributing to the CWE.

SUMMARY AND LEARNING POINTS

Riparian communities are found in nearly every geologic formation, geomorphic setting, and climatic regime. As a consequence, the hydrology, vegetation, and soils of a riparian community are both spatially and temporally variable. A stable stream channel and high-quality water are usually sustained in a healthy riparian community. However, riparian areas and their surrounding watersheds are continually exposed to a diversity of disturbances and have often been damaged as a result with the implementation of remedial treatments required. At the end of this chapter, you should be able to

(1) Explain the processes by which a healthy riparian system can sustain the hydrologic and water-quality benefits of a watershed.
(2) Understand the roles of riparian buffer strips in protecting streams and lakes.

(3) Describe the role of large-woody materials in stream dynamics and know the source of large-woody materials in and adjacent to the stream channel.
(4) Understand the importance of hydrologic and geologic settings when planning and implementing riparian rehabilitation treatments.
(5) Explain why it is important to know instream flow requirements in managing riparian communities.

Wetlands are important biological and hydrologic features of watersheds. In addition to supporting unique plant and animal communities, wetlands can reduce streamflow volumes and peak discharges to receiving stream channels through their storage function. They can also reduce the sediment delivery to receiving waters and improve the quality of water that is discharged to streams and rivers. After reading this chapter, you should be able to

(1) Explain the hydrologic conditions necessary for wetland formation.
(2) Describe the surface water and groundwater flow pathways and conditions that can lead to the formation of different types of wetlands.
(3) Develop a water budget for a wetland that has a perched groundwater system and a wetland that is an integral part of a regional groundwater system.
(4) Explain how wetlands affect flooding.
(5) Describe the mechanisms by which wetlands affect water quality.
(6) Provide examples of how management practices on the watershed of a wetland can result in losses and gains of wetland ecosystems.

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CHAPTER 14 *Watershed Management Issues*

INTRODUCTION

The previous two chapters focused on the hydrologic and water-quality effects of management practices, fire, and other land-use activities on wildland watersheds. With expanding pressures on wildland watersheds by a growing human population, many of the major issues of land and water deal with these pressures. Here, we will consider management issues associated with the effects of fragmentation of wildland watersheds caused by the encroachment of, and conversion to, agricultural croplands and urban development. Also, we discuss the means by which managers can proactively address the adverse effects of various land-use changes on the quantity and quality of flows from fragmented watersheds through integrated watershed management (IWM). Best Management Practices (BMPs) implemented to reduce or prevent excessive sedimentation and water pollution are discussed. The requirement to comply with laws and regulations established to maintain flows of high-quality water is also stressed. The possible impacts of climatic variability and change on water quantity, quality, and timing of streamflows are considered to IWM into a proper perspective. We also discuss the dilemma faced by many planners and managers of natural resources – that of coping with insufficient information to make informed decisions.

FRAGMENTATION OF WATERSHED LANDSCAPES

Fragmentation (parcelization) of ownerships and interests makes the management of watersheds a challenging task. One consequence of converting wildland watersheds to agricultural cropland or urban development has been and continues to be increased land fragmentation in both the public and private sectors (Alig and Plantinga, 2004; Alig et al., 2004). The fragmentation of privately owned lands on watersheds is often driven by economics where

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

Box 14.1

Fragmentation of Privately Owned Forests
Small mostly nonindustrial privately owned forests in the eastern USA and larger tracts of forest land owned by commercial timber companies throughout the country continue to be converted to agricultural croplands and urban development's alike. An average of 40.5 thousand ha of private forests have been fragmented annually by developments (Alig et al., 2004; Stein et al., 2005) with an additional 17.8 million ha expected to be developed for other land uses by 2030, representing more than 0.60 million ha each year. There is concern about how such changes in forest cover will affect the flow and quality of water and other ecosystem services. One consequence of this fragmentation has been to increase pressures placed on remaining contiguous forests located primarily in public forests for other ecosystem goods and services.

tax incentives are lacking to maintain or effectively manage these watersheds. Intensive land fragmentation in the United States has occurred with the conversion of small privately owned forests and land owned by commercial timber companies (Box 14.1).

Conversion to Agricultural Croplands

One of the largest conversions of wildland watersheds to agriculture croplands and to a lesser extent to urban developments in the world occurred in the north-central regions of the United States beginning in the late 1850s, reaching a peak in the late 1930s, and then continuing at a lower rate into the twenty-first century. An area larger than Minnesota, Wisconsin, and Michigan combined had been converted to mostly agricultural croplands and pastures by the middle of the 1990s (Verry, 2004). Many of the original forests in the eastern and southeastern regions of the country have also been converted to agriculture, although some areas have reverted back to forest conditions in recent decades. We discussed the impacts of these conversions in terms of the loss of wetlands (see Chapter 13). However, changes to watersheds and their hydrologic behavior extend beyond that of wetland loss alone.

The hydrologic impacts of converting wildland watersheds to agricultural croplands are determined by factors such as:

- the extent of conversion;
- whether rain-fed or irrigated agriculture is practiced;
- the type of original vegetative cover and type of crops to be grown;
- particularly whether they are annual or perennial crops;
- the season that the crops are grown;
- the cropping practices;
- the erodibility of the soils;

TABLE 14.1. Expected changes in hydrologic processes and responses resulting from the conversion of wildland watersheds (forests, grasslands) to agricultural croplands

Hydrologic process	Expected change[a]	Comment
Precipitation input	Unchanged	
Interception	Decrease	Depends on leaf area of crops and whether annual or perennial crop
Net precipitation	Increase	Due to decrease in interception losses
Infiltration capacity	Decrease	Depends on soil, crop, cropping practice
Transpiration	Decrease or remain same	Depends on whether annual or perennial crop and rooting depth differences
Surface runoff	Increase	Depends on soil and cropping practice; could be reduced if artificial drainage is practiced
Groundwater recharge	Uncertain	Depends on surface-groundwater linkages
Peak stormflow	Increase or variable	Increases for events with recurrence intervals $<$20–30 years, depends on soil, crop, cropping practice
Baseflow	Increase or remain same	Depends on whether cropping accompanied by artificial drainage and seasonal transpiration changes
Annual streamflow	Increase or remains same	Depends on extent of reduction in interception and transpiration

[a] Effects are exacerbated where forest cover is converted to annual crops.

- whether artificial drainage is practiced; and
- the extent to which wetlands and riparian systems are impaired or destroyed.

Describing how the conversion of wildland watersheds to agricultural croplands might affect hydrologic processes is a difficult task because of the varying impacts that the above factors can exert on these processes. Nevertheless, we can estimate changes in hydrologic processes and their hydrologic effects that would generally be expected to occur when the above conversions are made (Table 14.1).

We will consider the effects that conversions to agricultural croplands can have on streamflow regimes, water quality, and riparian communities and wetlands in a more general context below.

Streamflow Volumes and Regimes

Conversions from forests to agricultural croplands would generally be expected to increase the volume of streamflow due primarily to a reduction in annual *ET* on a watershed as explained in Chapter 12. Changes in streamflow regimes can also be significantly altered when the conversions take place on the variable sources areas of a watershed. These source areas can shrink or expand in size with changing hydrologic conditions (Hewlett and Hibbert, 1967). The effects of such changes are manifested mostly by increases in stormflow events. With the exception of extreme storms, peak stormflows have nearly tripled (on average) at the height of agricultural conversions in the north-central regions of the United States in comparison to stormflows when the landscapes were heavily forested (Verry, 2004).

Because the interception of precipitation is generally less on agricultural croplands than in either forests or woodlands, more precipitation is likely to reach the soil surface to increase surface or subsurface flow. Elimination of the litter and other organic material accumulations on the soil surface that absorbs much of the precipitation and the consequent increase in overland flow of water on the agricultural lands contribute to increased stormflows. The decrease in vegetative densities on these lands generally reduces the time to peak of stormflows and, in doing so, can increase the magnitude of the peak flow.

Conversions from grasslands, shrublands or savannah systems to annual agricultural crops would not be expected to have as severe of an impact on streamflow regimes as forested conversions. Generally these vegetative types do not exert as much influence on processes of interception, annual *ET* and surface–subsurface flow regimes as forests. However, extensive changes on the landscape that affect storage and conveyance of water can have significant changes (Box 14.2).

The conversion to agricultural crops does not always cause changes in streamflow response, as can be the case where wildland vegetation is converted to woody or other perennial crops. For example, the conversion of rainforests to tea plantations in Kenya, East Africa, resulted in relatively little effect on surface runoff or streamflow discharge (Edwards and Blackie, 1981). Furthermore, the cultivation of small areas in evergreen forests only increased streamflow discharges by less than 10%.

Water Quality

There is generally an increase in surface soil erosion following the conversion of forests or perennial grasses to agricultural crops because of increased exposure of soils to rainfall impacts and higher rates of soil erosion, especially on steeper slopes. Much of this sediment can be delivered into stream channels via overland flow. For example, sediment yields (per storm event) at the peak of agricultural development in the north-central regions of the United States in the late 1920s had increased more than 5 times in comparison to the estimated yields of sediment in the 1870s and had increased another 2.5 times by the early 1990s (Fitzpatrick et al., 1999 cited by Verry, 2004). However, these increases were small when compared to the extensive row-crop agriculture in the north-central and central Midwest regions where sediment yields have increased by up to two orders of magnitude since presettlement times (Evans et al., 2000). More generally, Welsch (1991) estimated that agricultural croplands in the United States account for 38% of the sediment deposited in the stream channels of watersheds each year.

Nutrient concentrations in streamflows often increase following the conversion to agriculture mostly because of the application of fertilizers to increase crop production. Total particulate phosphorus (P) loads to streams often increase directly with the increases in sediment loading from crop fields. Similar to the situation encountered when applying nitrogen-based fertilizers in forest ecosystems, urea and ammonia levels might remain below levels of concern in most situations. NO_3 does occur at high concentrations in certain landscape settings like the upper Midwest where subsurface drainage has been developed over large areas (Magner et al., 2004). The higher applications of N applied in croplands would be expected to result in much higher loads of NO_3 to streams than that expected from wildland applications. Other nutrient constituents (including phosphorus and chloride discussed in Chapter 11) in commercial fertilizers are likely to contribute to

Box 14.2

Hydrologic Changes due to Conversion of the Prairie Pothole Region to Annual Agricultural Cropping in the Upper Midwest (USA)

ET – Water Flow Changes

Conversion from perennial prairie grasslands to annual crops can reduce *ET* largely through reductions in the interception by plant residues of the prairie ecosystems. Comparisons of water budgets of prairie and maize ecosystems in Wisconsin (Brye et al., 2000) indicate that:

1. Interception by prairie residues (plant litter) covering the soil surface was 477 mm over a growing season with 681 mm of rainfall.
2. Total drainage of prairie, no-tillage maize, and chisel-plow tillage of maize amounted to 199 mm, 563 mm, and 793 mm respectively, over a 2-year period.

The results indicate that (1) conversions from native prairie to maize would significantly increase water flow, and (2) the increase in water flow if converted to no-tillage maize systems would be less than for chisel-plow tillage.

Drainage of Prairie Potholes

Concurrent with the extensive conversions of prairie ecosystems to corn (maize) and soybean crops across the upper Midwest following European settlement, more than 80% of the prairie potholes were drained via artificial subsurface drainage with tiles and plastic pipes. The consequences to watersheds have been:

- loss of capacity to store snowmelt and rainfall, and
- increase in land areas that contribute to streamflow connecting former pothole basins to ditches and stream channels.

Hydrologic Changes

The cumulative effects of conversions and drainage have increased total streamflow amounts average annual peakflows (Miller, 1999) and levels of baseflow in rivers of the upper Midwest (Zucker and Brown, 1998; Zhang and Schilling, 2006; Schilling et al., 2008). Although increases in annual precipitation have occurred over some basins in recent decades, the observed increases in flow in these basins are disproportionate to increases in precipitation alone (Lenhart et al., 2011). Furthermore, increased flows in the Mississippi River over the past 50 years have been attributed to land-use changes that have not been caused by changes in climate and precipitation regimes (Raymond et al., 2008).

FIGURE 14.1. Eutrophic water resulting from excessive nutrients from agricultural fertilizers applied in croplands (Photograph by Mary Presnail)

increases in the loading of pollutants to streams originating on agricultural landscapes with impacts on water quality, including eutrophication of water bodies downstream (Fig. 14.1).

Riparian Corridors and Wetlands

Elimination or partial elimination of riparian vegetation, drainage of wetlands, and changes in channel morphology can frequently accompany the conversion of a wildland watershed to agricultural croplands. All of these activities can increase in stormflows and sedimentation following the conversion (Baker et al., 2004; Verry, 2004). Construction of drainage ditches and culverts often compound these problems. The drainage of wetlands results in the loss of the natural storage provided by these ecosystems; consequences include a loss of wetland vegetation and a loss of critical watershed hydrologic functions (see Chapter 13). However, where ditches have been established from decades past, some hydrologic and ecological features can be improved by implementing alternative ditch designs. One design that shows promise is the Two-Stage ditch (Kramer, 2011). The Two-Stage ditch provides a small floodplain inside of the overall ditch geometry; this feature can buffer both sediment and nutrients and provide improved habitat for fish (Fig. 14.2).

Buffer strips of trees, shrubs, and perennial herbaceous plants in the riparian corridors of agricultural cropland can mitigate some of the adverse hydrologic impacts of converting to agricultural croplands. For example, infiltration rates and storage of soil water can be increased while surface runoff is decreased by these buffer strips (Schultz et al., 1995;

Box 14.3

Water-Quality Impacts of Wetland Drainage and Vegetative Conversions in the Prairie Pothole Region of the Upper Midwest (USA)

Wetland drainage represents a loss in wetland functions that affect water quality of streams in the upper Midwest. The Minnesota River Basin (MRB) has experienced the land-use changes and hydrologic responses described in Box 14.2 resulting in many tributaries that have impaired water quality, for example:

1. Increased flows of nutrients from fertilized croplands through drain tiles export high levels of nitrate-N directly into stream channels; most subsurface tiles discharge directly into ditches or stream channels and essentially bypass any mitigation of nutrients by riparian vegetation (Magner et al., 2004).
2. With fewer wetlands and little interaction between high nitrate discharges and riparian vegetation, there is less denitrification and uptake of nitrate occurring – resulting in high nitrate loading to rivers that flow into the Mississippi River and ultimately to the Gulf of Mexico (see hypoxia websites).
3. The loss of wetland storage has likely increased sediment export from watersheds during large storm events.
4. Increases in flows (see Box 14.2) have led to:
   - increased stream-channel erosion and greater instability of stream channels; and
   - increased levels of suspended sediment from streambanks and near-channel erosion that have resulted in high levels of turbidity (Lenhart et al., 2011).

As a result, many tributaries of the MRB are listed as impaired for nutrients and turbidity by the Minnesota Pollution Control Agency. The list can be accessed at the following web site: http://www.pca.state.mn.us/index.php/water/water-types-and-programs/minnesotas-impaired-waters-and-tmdls/assessment-and-listing/303d-list-of-impaired-waters.html (accessed January 30, 2012).

As a result of these listings, mitigation strategies and efforts are underway which include changes in cropping practices, rehabilitation measures for stream channels and riparian areas, and restoration of wetlands.

FIGURE 14.2. Two-stage ditch in southern Minnesota, USA, provides a buffer from cultivated land and roads (Photograph by Geoff Kramer)

Anderson et al., 2005). Streambanks are also better stabilized by the reduction in surface runoff that in turn reduces the amount of sediment and other pollutants delivered to the stream channels. Fast-growing tree species such as willow and poplar are generally preferred in the composition of buffer strips so that beneficial functioning of the strips can be established in the shortest period of time possible.

Wetlands that are drained in the conversions to either agricultural croplands or urban developments are in some instances in the United States required to be replaced with artificially constructed wetlands with the costs of this construction often incurred by the landowner.

Returning Agricultural Lands to Perennial Vegetation

Returning annual croplands to perennial vegetation (trees, shrubs, and herbaceous plants) has promise as methods of restoring the hydrologic functions that are associated with wildland watersheds. The effects of the Conservation Reserve Program (CRP) on water resources and the environment is one example of what these “reverse conversions” might be (Box 14.4). The CRP was established by the US Department of Agriculture (USDA) in 1985 as a voluntary program to reduce the soil loss on nearly 18 million ha of “highly erodible” agricultural croplands by planting trees, shrubs, and perennial herbaceous plants. In addition to the goal of reducing soil erosion, the CRP was instituted to improve environmental quality and provide income to farmers (Fig. 14.3). Enrolled farmers are compensated for one-half

Box 14.4

Impacts of the Conservation Reserve Program (CRP) on Water Resources and the Environment

Studies of the impacts of CRP have indicated significant reductions surface erosion and sediment loading in streams system. Sediment responses can be delayed, however, as near stream and instream sediments are remobilized (Davie and Lant, 1994). Depending on the location, other water-quality services of the program have been the abatement of nonpoint source (NPS) pollution, the "natural purification" of water, and the dilution of wastewater (Loomis et al., 2000). Among the unplanned ecological benefits of CRP, include a reduction in landscape fragmentation, increased fish and wildlife habitat, and an increase or maintenance of biological diversity.

of the costs of establishing these covers in exchange for retiring their lands for a 10-year period of time.

Once CRP lands are eligible for release, farmers can continue agricultural crop production, partially continue crop production, or retain the tree, shrub, and herbaceous covers. However, owners and managers of CRP lands need to monitor the impacts of their decisions on hydrology and especially water quality regardless of the decisions made.

FIGURE 14.3. CRP land croplands of southern Minnesota, USA. Perennial cover serves as wildlife habitat and functions to reduce nutrient and sediment loading to streams from adjacent croplands (Photograph by Mary Presnail)

Conversion to Urban Developments

Urbanization has been a "significant" cause of the fragmentation of forests and woodlands since the 1950s and it is anticipated that continuing urbanization will account for additional losses of these landscapes into the twenty-first century. Causes of this fragmentation are numerous and include the expansion of recreational developments, construction of private homes in remote forest environments, the expansion of suburban areas in forests near cities, and the associated roads and infrastructure of urbanization. The National Resource Inventory of the USDA indicated that there was a 34% increase in land converted from forests, woodlands, agricultural croplands, and open spaces to urban and other developed areas between 1982 and 1997 in the United States, with projections that such conversions would almost double from 1997 to 2025 (Alig et al., 2004). Urban developments in the United States are expected to grow the fastest in the western and southern regions.

Conversions of wildland systems to urban landscapes can have dramatic effects on hydrologic processes and the amount, timing and quality of water flows. Urban development effects vary depending on:

- the density of houses and other buildings;
- structural features of these buildings especially the type of roofs;
- the total area that is impervious, for example, streets, sidewalks, and parking lots;
- recontouring of land surfaces to accommodate the construction;
- alterations to stream channels such as realignment for road crossings and the associated changes to riparian vegetation and disruption to groundwater-surface water linkages in the riparian zone; and
- loss of riparian vegetation and wetlands.

Changes in the hydrologic processes following a conversion to urban developments are presented in Table 14.2. Keeping the factors impacting on these processes in mind precludes anything but general estimates, however.

TABLE 14.2. Changes in hydrologic processes and responses resulting from the conversion of wildland watersheds to urban developments

Hydrologic process	Type of change	Comment
Precipitation input	Unchanged	
Interception	Decrease	Due to reduced vegetative cover and increased impervious areas
Net precipitation	Increase	A function of interception change
Infiltration capacity	Decrease	Due to impervious layers created
Storage of water	Decrease	Related to decrease in infiltration
Transpiration	Decrease	Due to loss of vegetation
Surface runoff	Increase	Due to impervious layers created
Peak stormflow	Increase	Due to increase in surface runoff
Groundwater recharge	Decrease or unchanged	Depends on extent of impervious area in recharge zones
Baseflow – dry season flows	Decrease or no change	Depends on extent of impervious area and surface runoff vs. groundwater recharge over the landscape
Annual water yield	Increase	Due to reduced transpiration and interception and increased surface runoff

The cumulative effects of the impervious surfaces, the construction of storm-drainage systems, modifications to stream channels–including straightening, and the removal of the canopies of native vegetation can impact on the amount, timing, and suitability of people's water supplies. Conversely, the use of rain gardens in urban areas can promote infiltration and reduce surface runoff to storm sewer systems. Furthermore, some impervious surfaces in urban areas can be used as components of rainfall-harvesting systems that have promise for capturing water for further use.

Streamflow Regimes

The urbanization of wildland watersheds can alter the movement and storage of water as a consequence of the impervious areas created. Much of the change in streamflow results from the "paving" of the watersheds in developing urban areas (Finkenbine et al., 2000). These impervious areas reduce or almost eliminate the infiltration of rainfall, resulting in surface runoff that can significantly increase localized streamflow discharges (Box 14.5). Groundwater recharge is also largely reduced in response to increases in impervious area in a watershed.

Flashy flows of surface runoff occur from urban areas following high-intensity rainfall events. Streamflow resulting from the accelerated flows can exceed the bankfull stage of downstream channels and, as a consequence, lead to concurrent increases in flooding in some instances. The expansion of flood-prone areas and their frequent habitation by urban residents without adequate flood-protection measures can result in major economic losses.

Water Quality

The loss of vegetative cover, increases in construction and road building activities, degradation of riparian communities, and loss of wetlands that often accompany urban development can cause larger volumes of sediment-laden water to flow to a watershed outlet than occurred before urbanization. Sediment concentrations increase with increases in fine sediments from construction activities, the impervious surface generating even higher localized peak flows than agricultural conversions, and removal or partial removal of the buffering protection of riparian vegetation in mitigating the movement of soil particles in overland flows.

Box 14.5

Effects of Urbanization on Stormflow: An Example

Less than 10% of the rain falling on the watersheds of the Los Angeles River was converted into stormflow before 1930. However, almost 90% of this rainfall becomes stormflow since 1990 as a result of the unprecedented expansion of the Los Angeles metropolitan area onto these watersheds (Drennan et al., 2000). One of the consequences of this urbanization has been a reduction in the infiltration of rainfall caused largely by the impervious layers created by urban development.

Box 14.6

Monitoring Water Quality in a Municipality: One Example

Water for the residents of Tucson, Arizona, a metropolitan area of almost 1 million people, comes from groundwater pumped from 200 wells, a blend of recharged water delivered by the Central Arizona Project (CAP,) and groundwater from a recharge and recovery facility. The CAP transports water from the Colorado River to central Arizona through a system of canals and diversion ditches. Using the blended water has reduced groundwater pumping to where less than 50% of the total water supply for the city comes from pumping the groundwater aquifers. More than 20,000 individual tests of the quality of the drinking water are performed annually by the Division of Water and Operations. Constituents tested include sodium, mineral content, hardness, pH level, and temperature. The percentage of samples exceeding the primary standards for coliform bacteria and disinfectant chlorine established by the US Environmental Agency are also determined. The results of these tests are distributed monthly to people with their water bills by Tucson Water.

The quality of water can also become impaired by anthropogenic pollutants such as oils, salts, and other substances in the surface runoff from streets, sidewalks, and parking lots. Therefore, many municipalities monitor their water supplies on a continuing basis to insure the delivery of high-quality water to local inhabitants (Box 14.6). These monitoring programs are typically the responsibility of local water managers and organizations.

Riparian Corridors and Wetlands

Since ancient time, rivers and their riparian corridors have attracted people and, as a result, became sites for the establishment of large metropolitan communities and small villages. However, riparian areas and wetlands have not fared well in the process of converting wildland watersheds to urban areas. Protective riparian buffers are often partially or completely eliminated and wetlands are frequently drained in the construction of urbanized areas. The cumulative effects of such activities are often unwanted (Box 14.7).

A major problem facing the inhabitants of urban communities is finding an environmentally and economically acceptable solution to the often increased flooding of urban streams. Flood control measures can include stream channelization, which is, widening, straightening, or deepening a stream channel. However, channelization tends to adversely affect the physical and biological environment and reduce the aesthetic quality of the stream system. Levees and flood walls of concrete are constructed to help control flooding locally but people must consider their downstream effects, particularly where such levees and flood walls are not in place.

Box 14.7

Levees, Wetlands and Floodplains – An Example of Cumulative Effects

Riverine wetlands and floodplains with riparian vegetation have often become disconnected from rivers by the construction of levees. Levees are often constructed in urban areas to reduce flooding locally, only to increase streamflow discharges further downstream. At any one point in time and at any one location, the loss of functioning wetlands and floodplains might almost seem trivial. However, the cumulative effects of such losses along a river can have compounding effects on stormflow and flood peaks as suggested by Leopold (1994) in his post-flood analysis of the 1993 flood on the Upper Mississippi River, USA. The losses of storage along the river are additive to other land-use changes brought about by urbanization that can lead to increases in the velocity of streamflow, stormflow peaks, and in the stages (elevations) of rivers that correspond to stormflow peaks. Changes in channel sediment loads and other water-quality constituents can also be detrimental, as mentioned previously. Reversing these effects is not easy and likely involves improved management of urban watersheds but also a river basin perspective and an IWM plan that involves many different political units. The economics of mitigation interventions that involve restoration or rehabilitation of wetlands, riparian areas, and stream channels to correct such problems are prohibitive for most local communities.

Responses to Hydrologic Impacts Caused by Urbanization

In contrast to the hydrologic changes associated with timber harvesting and wildfire that are largely transitory because of forest vegetation regrowth following these disturbances, urban development is more permanent. Fortunately, low-impact methods of urban development and a host of BMPs can be implemented to prevent or mitigate these adverse impacts (see later section). Minimizing the loss of forests, woodlands, rangelands, and the adverse effects of urbanization on riparian areas, stream channels, and wetlands should, therefore, be incorporated into IWM programs to the extent possible. In addition, there is often a need for increasing local water supplies in areas of expanding human and livestock populations. An option for increasing local water supplies that can be integrated into IWM is developing water-harvesting systems.

WATER HARVESTING

Water harvesting is an ancient technique of developing water resources to augment the quantity and quality of existing water supplies in rural and urban areas where other sources are not available or too costly. *Water-harvesting systems* are methods for collecting and storing precipitation and resulting overland flows of water until it can be used for small-scale agricultural production or water for livestock, wildlife or domestic use. The systems require a catchment area to facilitate overland flows of water and a storage facility for the harvested water unless the water is to be directly applied to soil-plant systems. A water-distribution scheme is also required with those systems for irrigation. Water harvesting can be applied in almost any region with at least 100 mm of annual rainfall; however, sustainable farming requires annual rainfall of 250 mm or more (requiring adequate storage facilities).

Water-Harvesting Systems

The type of water-harvesting system depends on the topography, the type of treatment (if any) to increase runoff efficiency of the catchment, and the intended use of the harvested water (Frasier and Myers, 1983). Microcatchments, strip harvesting, and harvesting aprons are among the more common types.

Microcatchments are best suited for situations in which drought-resistant perennial plant species are grown (Renner and Frasier, 1995). The collection area can range from 10 to 1000 m^2 depending on the rainfall in the area and plant requirements. Plants are usually grown on the downslope side. Strip farming is a modification of the microcatchment method. Berms are erected on the contours with the area between them prepared as the collection area. Surface runoff between the berms is then concentrated above the downslope berm to irrigate the vegetation planted there. Only drought-hardy plants are grown with this system.

Apron-type systems are used primarily for livestock, wildlife, and domestic water supplies (Fig. 14.4). The catchment area (apron) is treated to increase runoff efficiency

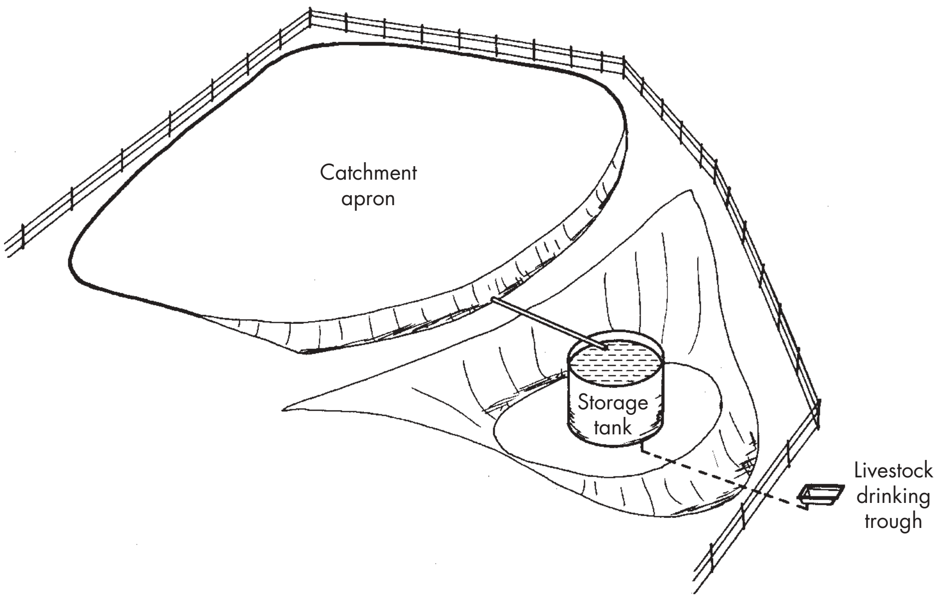

FIGURE 14.4. An example of a simple water-harvesting system (from Frasier and Myers, 1983)

unless an impermeable surface exists (Frasier, 1975). Gravel-covered asphalt-impregnated fiberglass is a common treatment. A storage tank is available with the necessary pipes and valves to conduct the water to drinking troughs or homes. The apron type is the simplest water-harvesting system to design. An approximation for the size of apron required can be obtained with the following equation:

$$A = b\frac{U}{P} \tag{14.1}$$

where A is the catchment area (m^2); b is equal to 1.13, which is a constant; U is the annual water requirement (L), and P is the average annual precipitation (mm).

Compartmental reservoirs of three storage areas are frequently recommended for agricultural production (Reig et al., 1988). Pumps are usually necessary to transfer water between ponds and drive the irrigation system. Gravity systems might be possible in steep terrain. Since relatively large quantities of irrigation water are usually required, treatment of catchments to increase runoff efficiency is often necessary. Sodium salt (NaCl) is effective where the soil has a sufficient quantity (about 10% or more) of expanding clays.

Catchment Areas

The catchment area should be impermeable to water to generate runoff. Examples of effective catchment surfaces include natural rock outcrops; smooth surfaces cleared of vegetation and rocks; surfaces treated with NaCl, silicones, latex, or oils; surfaces covered with concrete, butyl rubber, metal foil or plastic; and the rooftops of buildings. Runoff water must be nontoxic to humans, plants, and animals. The surface material should also be resistant to damage by hail or intense rainfall, wind, occasional animal traffic, moderate water flows, and plant growth. It should require minimum site preparation and simple maintenance.

Water Storage Facility

Among the more common types of storage facilities are the soil profile itself, excavated ponds, and tanks or cistern containers. Ancient water-harvesting systems were arrangements where runoff water was directed from hillsides onto cultivated areas to immediately store water in the soil for plant use. A problem with this arrangement was that insufficient water could be stored to offset a prolonged drought. Nevertheless, this method is still in use today to grow drought-resistant varieties of trees and other economic plants.

Excavated ponds are an economical means of storing the large quantities of water needed for agricultural production. However, evaporation and seepage are potential problems. Unfortunately, a simple, effective, economical method of evaporation suppression has not been developed. Seepage can account for 60–85% of the annual water loss. Chemical dispersing agents, bentonite, membrane liners, NaCl, and simple compaction can successfully sealed impoundments. Tanks or cisterns can be used effectively for livestock watering and domestic supplies. Seepage and evaporation are less difficult to control with these storage facilities. Any container capable of holding water can be a water storage facility. External water storage, a necessary component of a drinking-water supply system, can also be a part of a runoff-farming system, where some form of irrigation system applies the harvested water to the cropped area. The storage and water distribution facility is often the most expensive item in many water-harvesting systems.

Feasibility of Water Harvesting

Water collected from a catchment can contain organisms and water-soluble impurities from windblown dust deposited on the surface; chemical pollutants directly from the treatment (NaCl, silicone, tars, or oils); and weathering byproducts created by deterioration of the treatment materials. The quality of water from most surface treatments is usually adequate for livestock and wildlife but filters are needed in most cases if the water is for human consumption (Frasier, 1988). Most surface treatments that are most often used do not appear to adversely affect plant production.

Water-harvesting systems must include adequate storage to supply the amount of water that is needed at the time it is needed. For example, the need for livestock water depends largely on the grazing systems employed and the monthly distribution of rainfall. Domestic water supplies for people in the United States are 20–40 L/day for cooking, drinking, and washing. Water-harvesting systems for agriculture are more difficult to design. Little information might be available on the minimum water requirements of specific agricultural crops, although such data can be available for crops abundantly supplied with water. Equally important to the total water requirement is the timing of water needs. Therefore, the water-harvesting system must satisfy the seasonal pattern of use from establishment of the crops to its harvest. This information must be matched with the water supply to determine the frequency and amount of irrigation.

The feasibility of developing a water-harvesting system should be evaluated in the context of other alternatives to increase water such as untapped springs, a shallow groundwater table, or perched water that could be tapped with horizontal wells. All sources of water need to be evaluated with respect to costs, location, yield, dependability, and quality before embarking on the construction of a system.

Best Management Practices

The control of NPS pollution has focused on implementation of BMPs since passage of the Clean Water Act (CWA) by the Congress of the United States. The basis of the CWA was passed originally in 1948, with amendments in 1972 and 1977 that became the current form of the Act with the stated purpose of regulating the discharge of pollutants into waters of the United States and regulating the associated water-quality standards (see later).

BMPs are methods, measures, or practices that mitigate the movement of sediment, nutrient, pesticide, and other pollutants into stream systems by implementing varying nonstructural (vegetative) and structural (engineering) controls (Brown et al., 1993; Aust and Blinn, 2004; Ice et al., 2010). BMPs are applied to reduce or eliminate the delivery of the pollutants into receiving waters.

Adoption and application of BMPs help a watershed manager to achieve one or more of the following goals:

- maintain the integrity of stream systems;
- reduce surface runoff originating from areas of disturbance flowing into streams;
- minimize the movement of sediment and pollutants to surface-water and groundwater aquifers; and
- stabilize exposed soil surfaces through natural or artificial revegetation of disturbed sites.

BMPs in the United States can be either mandatory or voluntary depending on the state in question (Ice et al., 2010). The first authorized forest practices act to specifically address water-quality protection was adopted by Oregon in 1971 and, therefore, this action predated passage of the current form of the CWA. West Virginia and Kentucky have enacted mandatory BMPs regulations while Virginia has voluntary BMPs regulations except for activities associated with road construction and maintenance that are mandatory (Edwards and Willard, 2010). Other states have similar legislation.

Synthesis of BMPs

BMPs are often negotiated compromises among parties with interests in environmental protection (NRC, 2008). An iterative process of specification, application, and monitoring is the key to cost-effective BMPs use. One result of monitoring can be "fine-tuning" BMPs for future applications. A practical approach to this process is to initially prescribe BMPs that can be carefully studied to determine if they will control NPS pollution to meet water-quality standards and then reassess the array of BMPs as new information becomes available; this approach is referred to as *adaptive management*. The nature of NPS pollution requires iterative review across a wide array of landscapes and land use (Magner, 2011). However, relevant information about the overall cost-effectiveness of BMPs programs is often lacking. Magner and Brooks (2008) proposed the idea of a sentinel watershed to address this concern. Sentinel watersheds imply long-term study such that lessons learned can be inferred to similar land use superimposed upon similar climatic and geologic regimes.

BMPs to mitigate the erosion–sedimentation–nutrient processes are known for many silvicultural treatments, livestock-grazing practices, road-related soil disturbances, and agricultural activities (Moore et al., 1979; Lynch et al., 1985; Chaney et al., 1990). One example of a BMP in forest management is retaining riparian buffer strips to reduce the flow of sediment and other forms of NPS pollution into streams. Implementing a set of BMPs to protect water quality and wetlands is another example (Box 14.8). However, BMPs to control some types of pollutants are incomplete or unknown.

Road construction can be modified to enhance the restoration of sites previously damaged or impaired. For example, BMP techniques for the recovery of wet meadows that might be impacted by roads include relocation of site-damaging road segments; modification or replacement of channel-crossing structures by culverts, fords, or bridges; and the modification of ditch systems and cross drains to disperse streamflows (LaFayette et al., 1992). The structural (engineering) objective in this case becomes one of using road-drainage structures to conserve available surface and subsurface flows and nurture native vegetation to sustain ecosystem integrity.

Effectiveness of BMPs

Monitoring efforts reveal that the use and effectiveness of BMPs is increasing and that their implementation generally results in compliance with water-quality standards. A multitude of studies have compared sediment and nutrient loadings from undisturbed watersheds versus treatment watersheds on which BMPs have been used to estimate the effectiveness of water-quality compliance. The findings of these studies have been relatively consistent, showing that undesirable effects of land-use practices can be avoided or minimized when BMPs are properly implemented (Fig. 14.5). For example, studies in the eastern USA indicate the BMPs have significantly lowered NPS pollution from areas where trees have

Box 14.8

Best Management Practices: Forest Management for Protecting Water Quality and Wetlands (MDNR, 1995)
A set of BMPs developed in Minnesota for forest resource management targeted the following:

- managing fuel, lubricants and equipment – practices include designating the least hazardous areas for fueling and maintaining vehicles and equipment, reporting procedures for spills, etc.;
- providing filter strips of vegetation adjacent to lakes, streams and open-water wetlands to reduce sediment, nutrient, and pesticide runoff from entering the respective water body – practices specify widths of filter strips based on the slope of the land and limitations of soil exposure and disturbance within filter strips;
- provision of shade strips of vegetation adjacent to lakes, streams and open-water wetlands to moderate water temperature – specific requirements of 100 ft. (35 m) are designed for trout streams (cold-water species), required minimum densities of trees for different management objectives, and recommendations for minimizing disturbance;
- wetland protection measures – stated requirements for planning activities and for minimizing disturbance of wetland sites during forest management activities;
- building and maintaining forest roads – detailed and comprehensive set of guidelines for planning roads, selecting types of roads, water-crossing specifications, drainage requirements, etc.
- timber harvesting guidelines – recommendations for developing a plan, designing the harvesting site to minimize runoff and disturbance to water bodies, and postharvesting recommendations;
- mechanical site preparation guidelines – planning, design, and operational recommendations;
- pesticide use guidelines – incorporating integrated peat management strategies, selecting pesticides, measures to minimize contamination potential, and managing all phases of pesticide use; and
- prescribed burning guidelines – planning considerations and practices for eliminating the hazard of wildfires, facilitating revegetation, improving wildlife habitat, controlling insects and diseases, and reducing potential for erosion and sediment deposition into water bodies.

FIGURE 14.5. Undesirable effects on soils and water such as illustrated in this photo can be avoided when BMPs are properly implemented. In this case, a perennial vegetated buffer would provide some degree of protection for the water quality in this stream (Photograph by Mary Presnail) (For a color version of this photo, see the color plate section)

been clearcut and roads constructed (Vowell, 2001; Aust and Blinn, 2004). Studies have also shown that while this effectiveness is site dependent (Blinn and Kilgore, 2001; Lee et al., 2004), generalizations are possible. For example, the effectiveness of BMPs such as establishing or retaining riparian buffer strips to reduce the concentrations of sediment and other forms of NPS pollutants in streams is noteworthy (Ice, 2004).

To gain a better understanding of the effectiveness of BMPs in eastern regions of the United States, Edwards and Willard (2010) compared BMP efficiencies from paired-watershed studies in which timber was harvested on one watershed with implementation of BMPs while timber on the second watershed was harvested without BMPs. *Efficiency* was measured as the percent of pollutant reduction (sediment or nutrients) achieved by applying

BMPs. For each of the studies, percent efficiencies of a BMP for a specified time period (E) were calculated from:

$$E = \frac{BMP_o - BMP_i}{BMP_o} \tag{14.2}$$

where BMP_o is the sediment or nutrient loading from a watershed on which a BMP was not implemented; and BMP_i is the nutrient loading from a watershed on which a BMP was implemented.

Edwards and Willard (2010) reported that the efficiencies of BMPs in reducing sediment yields ranged from about 55 to 95% up to 1 year after timber harvesting. For nutrients, BMP efficiencies were higher for total nitrogen (60–80%) and phosphorus (85%) in both particulate and sediment-bound forms than for nitrate nitrogen (<15%) that occurs primarily in the dissolved phase. It was also concluded that the BMPs were more effective in reducing pollutants associated with surface water than with subsurface flow.

Further Information on BMPs

It is suggested that the reader access the Watershed Academy's module on BMPs for further information on BMPs for watershed management. In addition to providing general guidance on BMPs for minimizing the impacts of NPS pollution, this module contains a training course prepared by the US Forest Service and the Environmental Protection Agency on implementing forest management practices for the control of NPS pollution. The Watershed Academy website is found at: http://www.epa.gov/owow/watershed/wacademy/acad2000/forestry/right1.htm (accessed January 30, 2012).

Regulatory Compliance

Watershed management practices must be planned and implemented to comply with laws and regulations pertaining to the maintenance of high-quality flows of water from upland watersheds regardless of the management purpose. Regulatory compliance in the United States is required at both federal and state levels.

Federal Regulations

Among the federal regulations confronting a watershed manager in the United States are the National Environmental Act, the CWA, and a host of other federal laws pertaining to implementing a watershed management practice. The requirements of some of these regulations are summarized below.

National Environmental Policy Act

The National Environmental Policy Act, commonly known simply as NEPA, specifies the requirements that all federal agencies must follow to promote the enhancement of environmental quality. This Act also established the Council on Environmental Quality (CEQ) to oversee the coordination of the NEPA process among the agencies. Watershed management practices on public lands are subject to the NEPA provisions, and, therefore, a watershed manager of these lands must be familiar with the NEPA requirements. However, the Act does not apply to purely state or private actions.

The NEPA Process. It is necessary that a watershed manager of public lands knows the NEPA process, recognizes the obligations in implementing its provisions, and complies with its requirements. The NEPA process is initiated when an agency initiates a proposal to take action; for example, to implement a riparian-stream channel rehabilitation project. Once a determination has been made that the proposed action is within the NEPA provisions, there are three sequential levels of analysis that the responsible agency is required to consider in complying with the law: (1) preparation of a Categorical Exclusion (CE), (2) preparation of an Environmental Assessment (EA) and a Finding of No Significant Impact (FONSI), and (3) preparation of an Environmental Impact Statement (EIS).

A CE is a category of actions that the agency has determined does not individually or cumulatively exert a significant impact on the quality of the environment. If a proposed action is included in the description for a listed CE by the implementing agency, that agency must then determine that "no extraordinary circumstances" exist that might cause the action to have a significant effect in the particular situation.

An EA is a "screening document" describing the environmental effects of the proposed action and alternative means to achieve its objectives. The purpose of an EA is to:

- provide sufficient evidence and analysis to determine whether to prepare an EIS;
- to facilitate an agency's compliance with NEPA when preparation of an EIS is deemed not necessary; and
- facilitate the preparation of an EIS when one is necessary.

If preparation of an EIS is not necessary, the agency might produce a FONSI that outlines the reasons why the proposed action will not have a significant effect on the environment. It must also include the EA or a summary of the EA that supports the (FONSI) determination. If it is determined that the proposed action does not fall within a designated CE or does not qualify for a FONSI, the responsible agency or agencies must prepare an EIS. The purpose of an EIS is to help officials make informed decisions on the environmental impacts of the proposed action and the alternatives available. An EIS is required to describe:

- the environmental consequences of the proposed action;
- adverse environmental impacts that cannot be avoided should the proposal be implemented;
- reasonable alternatives to the proposed action;
- the relationship between local short-term uses and impacts on the environment and the maintenance of long-term productivity; and
- irreversible and irretrievable commitments of resources involved in the proposed action should it be implemented.

Federal Agency Role. The agency that will carry out the proposed action is responsible for complying with the NEPA requirements. A *lead agency* supervises the environmental analysis. This agency might work with state, tribal, or local agencies as joint federal agencies. A *cooperating agency* assists the lead agency in the scoping process, preparing the environmental analysis, and providing staff support upon request.

Agencies can refer to CEQ on disagreements concerning a proposed action that might cause "unsatisfactory" environmental effects. CEQ's role is to develop findings and recommendations consistent with the goals of NEPA. The Environmental Protection Agency

(EPA), like other federal agencies, prepares and reviews NEPA documents. However, EPA also has responsibilities to review and publicly comment on the environmental impacts of most of actions that are subject of EISs. EPA might be required to refer the matter to CEQ if it determines that the action is environmentally unsatisfactory.

The Public's Role. The public has a role in the NEPA process in providing input on what issues should be addressed in an EIS and commenting on the finding in NEPA documents provided by the lead agency. The public can also participate in the NEPA process by attending NEPA-related hearings or public meetings and submitting comments on the proposed action to the designated lead agency. The lead agency must take into consideration all of the comments received from the public and other parties on the NEPA documents in the specified comment period.

Clean Water Act

The CWA is the primary law in the United States that addresses protection of the quality of rivers and streams, lakes, coastal areas, and to lesser extent than the Safe Drinking Water Act (SDWA), groundwater aquifers. A primary goal of the CWA is restoring impaired waters and ensuring that the implementation of a planned action will maintain or improve the physical, chemical, and biological integrity of these bodies of water. Section 303(d) or the CWA was not fully utilized for several decades because the EPA's focus was point-source (PS) pollution compliance. A series of legal actions helped states understand the 303(d) list could be used to spotlight NPS pollution. For example, the total maximum daily load (TMDL) of total sediments entering into a stream system can be established by the regulations specified in this law (Box 14.9). Because the total sediment load at any point in the stream is a function of everything that influences the erosion and sedimentation above the point, that individual and collective influences of these upstream land uses must be addressed to meet the state water-quality standards. However, TMDL regulations only mitigate a problem after it has occurred.

The EPA, the agency responsible for "enforcing" the CWA, initially placed emphasis on the more traceable problems of PS pollution. However, the agency later recognized that water-quality degradation from NPS pollution as another barrier to meeting established water-quality goals. Section 208 of the Act specifically addresses NPS pollution and designates timber harvesting, silvicultural cuttings, livestock grazing, and road construction activities as significant potential of pollution (Brown et al., 1993). However, section 208 ceased to be helpful when funding dried-up; today NPS funding comes through section 319 of the CWA. A framework of standards, technical tools, and financial assistance is provided by CWA to address the causes of PS and NPS pollution and the resulting poor water quality. These causes can be:

- streamflow originating on watersheds in an environmentally unsatisfactory condition;
- polluted runoff from rural and urban areas; and
- municipal- and industrial-wastewater discharges.

Among other provisions, the CWA authorized a program of grants to help build publicly owned wastewater-treatment plants which then transitioned to a revolving-loan program. Section 319 and portions of the Farm Bill provide funding to rural landowners to implement

Box 14.9

A General Procedure for the Preparation of a TMDL

The problem of maintaining water quality can be compounded by the need to develop a TMDL for a water body found to have water-quality conditions that limit its ability to meet its designated beneficial uses as mandated by the CWA. A TMDL for a water body is the sum of the waste-load allocations (WLA) from identified point sources and the sum of the load allocations (LA) from nonpoint pollution sources and a margin of safety (MOS) within the water body's watershed. In some areas, an allocation for future growth must be added to the TMDL equation. A TMDL for the water body should identify current WLA + LA + MOS to the water body. With this information, allocations to the known point and nonpoint pollution sources within the water body's watershed are defined by tools like flow and load duration curves or models. However, preparation of a TMDL is a difficult task largely because of the problem of understanding and analyzing the mode of conveyance of diffuse NPSs. More conventional hydraulic methods can often be used for monitoring and analyzing the discharge of pollutants from point sources such as through a pipe. Identifying nonpoint pollutants is a more difficult task and requires not only monitoring and modeling but also stakeholder communication. Both nonpoint and point sources of pollution must be included in a complete TMDL.

BMPs for control NPS pollution. BMPs are considered to represent the most practical approach to controlling pollution and securing the flows of clean water from watershed landscapes.

Other Federal Regulations

The SDWA, the Endangered Species Act (ESA), the Clean Air Act (CAA), and the Federal Insecticide, Fungicide, and Rodenticide Act (FIFRA) are among the other federal regulations that can pertain to planning and implementing a watershed management practice.

Safe Drinking Water Act. The SDWA was established to protect the quality of drinking water in the United States. This law focuses on all waters designated for drinking whether the source is above ground or underground. SDWA authorizes EPA to establish minimum water-quality standards to protect tap water and requires owners or operators of public water systems to comply with primary (health-related) standards. State governments that are approved to implement the standards for EPA are encouraged to develop secondary (nuisance-related) standards. Under the provisions of SDWA, EPA also establishes minimum standards for state programs to protect underground sources of drinking water from endangerment by the underground injection of potential pollutants.

Endangered Species Act. The ESA is an overriding consideration in the planning and managing of almost all natural resources in the United States. This Act requires the conservation of threatened and endangered plants and animals and the habitats in which they are found. The lead agencies for implementing ESA are the US Fish and Wildlife Service (FWS) and the National Oceanic and the Fisheries Service of the Atmospheric Administration (NOAA). The FWS maintains a list of endangered trees, shrubs, herbaceous plants, mammals, reptiles, birds, insects, fish, and crustaceans. ESA requires the lead agencies in consultation with FWS and NOAA to ensure that a proposed action is not likely to jeopardize the continued existence of any listed species or result in the destruction or adverse modification of the critical habitat of the species. The Act also prohibits any action that causes a "taking" of a listed species for any purpose.

Clean Air Act. The CAA is a law that regulates air emissions from stationary or mobile sources. Among other things, this law authorizes EPA to establish National Ambient Air Quality Standards (NAAQS) to protect public health and public welfare and regulate emissions of air pollutants. As such, CAA comes into play when a watershed manager proposes the use of prescribed burning as a management tool. In preparing the prescription for this action, the manager must consider NAAQS at the federal level and state regulations that might also apply. A prescribed-burning treatment must be delayed or even cancelled when the emission standards specified by these regulations cannot be met because of the resulting smoke.

Federal Insecticide, Fungicide, and Rodenticide Act. The FIFRA provides for the regulation of pesticide distribution, sale, and use. All pesticides to be used in the United States must be registered (licensed) by EPA to prevent unreasonable adverse effects on the environment. The term "unreasonable" is defined by FIFRA to include any unreasonable risk to people or the environment taking into account the economic, social, and environmental costs and benefits of the use of any pesticide. This requirement means that only registered pesticides can be used by a watershed manager to convert one vegetation type to another type to alter streamflow regimes or following a wildfire to control the regrowth of unwanted vegetation.

A further discussion of the federal regulations that are relevant to planning and implementing a proposed watershed management practice is not a purpose of this book. The reader is referred to the internet and the appendix of the report on the hydrologic impacts of forest management published by the National Research Council (2008) for additional information on these regulations.

State Regulations

States in the United States also have laws and regulations that apply to watershed management practices. Most of these regulations are administered by a state agency with responsibilities to protect the public's interests specified by the state. Furthermore, because many of the federal regulations have become factors in the management of watersheds on state and private lands, in some cases, states have also been delegated the authority to administer some of the federal laws and regulations by the responsible federal agency.

Many states have established instream flow requirements to meet specified uses of water. While the specifics of these requirements vary from state-to-state, some or all of the uses that must be met include municipal drinking water, aquatic habitats, recreational purposes, maintaining riparian vegetation and wetlands, navigation, hydropower, and water quality including waste assimilation. Among the many states in the southeastern region with instream flow requirements are Georgia, North and South Carolina, and Tennessee. Other states with these requirements include Michigan, New Hampshire, Texas, and the State of Washington. The State of Washington also has regulations that address the needs of people who rely on out-of-stream uses of water.

State regulations that encourage a watershed as a planning basis are also emerging. However, the provisions of many of the state regulations and laws are voluntary and can call for pluralistic representation from state agencies, local government entities, representatives of the economic interests of people in the area of concern, and the general public. The main goal of these regulations is to collaboratively developing plans for the implementing the integrated management of land, water, and other natural resources on a state basis. The State of Minnesota took a major step toward watershed management in 2008 with the passage of a constitutional amendment directed at protecting and restoring land and water entitled Clean Water Land and Legacy. There are 81 major watersheds in Minnesota and the current plan is to systematically work around the state in a 10-year window to assess the overall ecological health of the watersheds. Land and water defined as high quality will be designated for more protection and water bodies that are identified as impaired will be restored. IWM in the form of local controls and BMPs will be implemented in specific areas defined as *priority management zones* (PMZs). The concept behind the PMZ is tailored targeting of limited funds to bring about maximum protection and/or restoration (Magner, 2011).

Climatic Variability

Climatic variability is one of the more challenging issues confronting watershed managers and, more generally, the people of the world at this time. Some people suggest that the climate is getting warmer (National Assessment Synthesis Team, 2008; National Research Council, 2010), some suggest that the climate is becoming drier (Overpeck and Udall, 2010), and still others believe that climatic changes are part of the natural variability of longer-term climatic conditions (Hulme, 2009). While there is increasing evidence that we are undergoing a warming trend across the globe, the extent to which this warming is human induced, an acceleration of natural warming trends, or embedded in the natural climatic variability and change has become a controversial topic among climatologists, natural resources managers, politicians, and concerned lay people. There are suggestions that global warming is causing precipitation to increase or intensify in some parts of the world while elsewhere precipitation is decreasing and possibly leading to more pronounced droughts. Given the uncertainty associated with these predictions, achieving the goals of IWM requires that people plan and manage land and water resources in ways that can be sustained under conditions of climatic variability and weather extremes.

Impacts on Water Resources

Climatic variability can influence the functioning of watersheds and the consequent changes in the volume, timing, and quality of water flowing from them. Model outputs that simulate watershed responses to climate change and changes in streamflow regime in some regions

FIGURE 14.6. Snowpacks on watersheds in colder climates could be reduced with the initiation of snowmelt runoff earlier in the season with increased global warming (Photograph by Mark Davidson)

of the world suggest a trend of declining water yields (Running and Nemani, 1991). For example, a simulation of the effects of climate change in the western USA indicated that a predicted warming of 1–2°C in the next half-century would result in a large reduction in mountain snowpack and a commensurate reduction in water storage (Barnett et al., 2004). As a consequence, snowpack amounts and persistence on watersheds in colder climates would be reduced with the initiation of snowmelt runoff will likely shift to earlier in the season with spring peak runoff occurring up to 3 weeks earlier in some regions (Fig. 14.6).

The future demands placed on water resources will likely not be met in many regions of the United States and other parts of the world, even with the most "conservative" estimates of climatic change. While water will continue to be a key issue in all regions into the future, changes in water resources attributed to climatic variability will vary regionally (Table 14.3). Therefore, regional perspective to future water resources is necessary because of the regional difference in climatic conditions (Table 14.3).

Planners and managers of natural resources should prepare a strategy to cope with changes in climatic conditions while sustaining natural resources for commodities and amenities. We suggest that it will be possible to sustain the benefits of IWM while coping with future climatic variability. Importantly, a flexible planning process is required to do so (see Chapter 15).

TABLE 14.3. Observed water-related changes in regions of the United States during the last century (modified from Karl et al., 2010)

Observed change	Direct of change	Region affected
Annual precipitation	Increasing	Most of the country
Annual precipitation	Decreasing	Southwest
Frequency of "large" precipitation events	Increasing	Most of the country
Proportion of precipitation falling as snow	Decreasing	West and Northeast
Snowpack water equivalent	Decreasing	Southwest
Duration and extent of snow cover	Decreasing	Most of the country
Surface runoff and streamflow	Decreasing	Colorado and Columbia River Basins
Streamflow	Increasing	Most of the East
Periods of drought	Increasing	Parts of West and East

Planning Protocols for Coping with Climatic Variability

A number of protocols are available to integrate climatic variability into the planning of watershed management within the uncertainty of future climates embedded in the process (Furniss et al., 2010). Included among these protocols are:

- adapting to climate variability by improving watershed resilience;
- advancing and sharing knowledge about climatic variability and the availability of water resources;
- integrating climatic variability into planning efforts; and
- implementing practices that protect and maintain or restore watershed processes and services.

Flexibility is paramount in coping with climatic variability because of the risks associated with forecasting future climates. Identifying the economic, environmental, and social costs associated with changing climatic conditions is also required. Recognizing the linkages among people, their natural resources, and the institutions responsible for planning and managing watershed resources is necessary in this process that must provide for:

- holistic considerations of the interactions among the uses of land, water, and other natural resources currently on a watershed and the expected uses into the future;
- a recognition of the risks of failure in achieving the goals and objectives originally specified in the planning process and possible alternatives if the risks cannot be overcome; and
- organizational capabilities and institutional arrangements necessary for managing the array of natural resources on a watershed basis for sustainable use in the long term.

Use of Hydrologic Simulation Models

Hydrologic simulation models that provide predictions of the future availability of high-quality water resources for a wide range of climatic conditions are often helpful in the planning process (see Chapter 16 for more details). Among the hydrologic changes embedded with most of these simulation models are rainfall and temperature regimes. Climate and associated vegetative changes will affect *ET* losses and, as a result, the availability of water resources. One such model with these capabilities is TOPMODEL (Beven and Kirkby, 1979) that simulates the responses of water resources to a sequence of time series representing rainfall and temperature regimes. These time series can be incorporated in a stochastic model when the simulation goal is forecasting the probability of future water resources within a framework of the climatic variability that might be encountered.

The time series of rainfall regimes obtained by applying TOPMODEL, the calibration of observed rainfall events against historical rainfall events, and a projection of this calibration to generate a long-term data set of future rainfall events are required in forecasting the impacts of climatic variability change on future water resources. However, TOPMODEL must be calibrated to the local heterogeneity and anisotropic conditions when an appropriate stochastic model is incorporated into the planning effort. Once these requirements are satisfied, the simulations of future rainfall events and water flows can be incorporated into the process to select the most suitable watershed management practice to implement.

Use of Stochastic Models

Stochastic models can play a role in selecting watershed management practices for implementation under conditions of climatic variability by providing a set of scenarios of climatic conditions that have an estimated probability of occurring in the future. More specifically, stochastic models estimate the probability distributions of future climates (outcomes) by allowing for random variation in one or more of the specified inputs over time. This random variation is derived from observed and projected fluctuations in historical climatic data for a specified period of time by applying time-series techniques. The distributions of these future outcomes are based on a large number of simulations called *stochastic projections* that reflect the random variation in the specified input(s).

Incorporated into stochastic models are statistical properties (means, standard errors, etc.) of the available data sets representing the climatic conditions of the watershed to be managed and the laws of probability that are the foundation for generating the sequences of data sets representing the probabilities of future climatic events (Duckstein et al., 1972; Fogel and Duckstein, 1982, and others). The climatic events with the highest probability of occurring in the future then become a basic input to the general planning process. Stochastic models can be applied to climatic problems such as estimating the probabilities that a sequence of selected increased rainfall events, warming temperature regimes, or drought conditions might occur in the future. Alterative courses of action should also be known if it is determined that the watershed management practice initially selected for implementation in the planning process is shown subsequently not to be a feasible choice or that it fails to meet its goals after its implementation.

The following scenario is an example of how stochastic models can be used to develop watershed management options. If a municipality was faced with the need to increase its water supplies, one option could be the development of water-harvesting systems. If a stochastic model indicated that the probability is too high that either the rainfall amounts or the sequence of rainfall events would not provide sufficient water at times when it is most needed, a lower-risk practice such as recycling waste water might be selected for implementation even though the goal of increasing water supplies was not fully satisfied. This process can be repeated as necessary until a compromise is reached between the occurrence of rainfall events and practices suitable for achieving the goal of increasing water supplies.

Further Information on Climatic Variability

The Climate Change Research Center (CCRC) of the US Forest Service offers information and tools to address climatic variability in the context of planning and implementing watershed management practices. The CCRC provides:

- updated information on current and future climatic trends;
- decision-support models to assist in the selection of a watershed management practice for implementation;
- maps including the *National Road Map for Responding to Climate Change*;
- simulations of estimated impacts of future climatic trends; and
- case studies.

The URL of the website for CCRC is http://www.fs.fed.us/climatechange/climate-update.shtml (accessed January 30, 2012).

INSUFFICIENT INFORMATION FOR DECISION MAKING

Another issue of possible concern in achieving the goals of IWM is the absence of long-term data sets for planning and preparing responses to hydrologic variability – such as droughts and floods. This issue centers on not having sufficient information on the hydrologic performance of a watershed or river basin to make informed decisions. While relevant short-term data might be available from some local community or associated with a short-term study, people may not know of its existence or how to access such data when its existence is known. Furthermore, by compiling short-term data sets over a region and for different time periods, an idea of certain characteristics such as climate or streamflow trends and variability may be gained.

Sources of Short-Term Information

A variety of hydrometeorological data have been collected on experimental watersheds for the period specified in the research plan and then the collection of these data has often been terminated. Examples of these studies were presented in Chapter 12 with some of the websites containing the data listed in the Webliography of the chapter. However, many of the streamflow records from these earlier studies have been lost because the storage media for data such as unpublished field measurements or computer-processing cards are outdated (Stednick et al., 2004).

While increasingly more meteorological, streamflow, and water-quality data are available on the internet as mentioned above, the lack of a standardized protocol for the deposition of these data in the United States and elsewhere remains a problem. According to a review of the data management needs in watershed management by Stednick et al. (2004), the US Geological Survey has established a protocol for its gaging stations but reporting of the data collected on forest, woodland, and rangeland watersheds do not necessarily follow this protocol. Water-quality data are also available from the EPA Storage and Retrieval System, the National Stream Water Quality Assessment Program, and the Long-Term Ecological Research Program of the National Science Foundation but not in consistent formats. The URLs for the websites of these respective data sources are listed in the Webliography at the end of this chapter.

Obtaining Long-Term Information

The lack of long-term data sets is even more critical to a watershed manager when confronted with having to make a decision on the future impacts of a proposed management practice or use of watershed resources. This dilemma becomes more acute when coping with the issue of climatic variability and the related hydrologic responses. Not having a long-term perspective can be alleviated to some extent by coupling the outputs of hydrologic simulation and stochastic models to generate scenarios of probabilities of the occurrence of meteorological and hydrologic conditions. These scenarios can then be considered in examining a particular watershed management practice and the responses to its implementation to determine if IWM objectives can be met. If it is then concluded that the risks are too high that the required conditions will not occur in the foreseeable future, a decision to replace the proposed watershed management practice with an alternative might be deemed appropriate.

One approach to obtaining a long-term (historical) perspective of the hydrologic functioning of a watershed or river basin is through dendrochronology or tree-ring analysis. *Dendrochronology* is a set of measurement techniques and analytical procedures by which the annual growth rings of sample trees are referenced to the corresponding years of their formation (Stokes and Smiley, 1968; Fritts; 1976; and others). A time-series chronology of the changes in the growing environment of the trees is then reconstructed from interpretations of the morphological or chemical properties of the growth rings.

The reconstruction of the effects of meteorological and hydrologic factors on the annual growth rings is correlated with the available (but often short-term) meteorological and hydrologic information. The resulting relationships between the two data sets are then extrapolated back into the past to generate a historical pattern of the meteorological and hydrologic information. The pattern of historical information obtained can then be projected to develop a pattern of future meteorological and hydrologic changes. An assumption that must be made in this extrapolation is that what has happened in the past will also occur in the future with a high level of certainty. This approach provides a more accurate perspective for decision making than was possible with only the relatively short-term information.

One application of dendrochronology in IWM is placing short-term streamflow records into a longer-term framework to determine whether the short-term streamflow record is representative of the historical streamflow in a watershed or river basin. Decisions based on the available streamflow record can be "rectified" if it is determined that the short-term record is not representative of the historical situation. An example of the use of dendrochronology in such a situation is presented in Box 14.10.

SUMMARY AND LEARNING POINTS

The hydrologic and water-quality effects of converting wildland watersheds to agricultural croplands or urban lands can be substantial. Fragmentation or the conversion of forest cover can in general cause greater changes in streamflow quantity, timing, and quality

Box 14.10

Dilemma of Making Informed Decision with Insufficient Information: An Example of the Application of Dendrochronology

One of the more noteworthy examples of the problem that can be encountered with a lack of long-term information on streamflow regimes resulted in an over-allocation of the water flowing in the Colorado River Basin of the western USA. Planners and decision-makers met in 1920 to allocate the rights-to-water flowing from the states in the upper basin into the states in the lower basin of the river. The planners and decision-makers estimated that the annual streamflow volume at the point of allocation averaged 19 985 billion m^3. This estimate was derived from the short-term available record of streamflow at the point of allocation that spanned the period of 1906 to the time that a decision had to be made, or approximately 14 years.

Stockton and Jacoby (1976) reconstructed the historical pattern of streamflow volumes at the point of allocation for the past 450 years with a time series developed from a chronology of annual tree rings to later place the available but relatively short-term record of streamflow volumes into a more representative long-term perspective. Their analysis showed that the streamflow volumes from the record available to the planners and decision-makers coincided with the longest period of sustained high streamflows in the 450-year period. Therefore, the record of streamflow volumes available in 1920 was not indicative of the historical streamflow regimes of the Colorado River and that the allocation of water into the lower basin had been based on an anomalously high record of streamflow volumes. As a consequence, the lower streamflow volumes more commonly encountered at the point of allocation has caused severe shortages of the available water when all of the states involved demanded their share of the water originally allocated to them.

than conversions to other types of wildland vegetation. The changing watershed landscape with expanding human populations presents challenges for achieving the goals of IWM. Disruptions in water supplies can often be offset locally with the development of water-harvesting systems. Restoration or rehabilitation of disturbed or degraded areas to predisturbance conditions may be desirable but not practical in most cases. BMPs are tools to mitigate NPS pollution to comply with regulations to help protect and sustain watershed resources. Minnesota has adopted a PMZ approach to strategically place BMPs on the landscape. However, long-term solutions will require *adaptive* management practices to cope with climatic variability. Doing so with the uncertainty of future climates and facing the dilemma of having insufficient information for informed decision making are among the issues confronted by watershed managers. After completing this chapter, therefore, the reader should be able to develop a list of IWM practices including BMPs to address hydrologic and water-quality problems due to fragmentation of perennial vegetative cover on watersheds. Specific questions to be addressed include the following:

1. What are the hydrologic impacts that can result from the fragmentation of wildland watersheds in their conversion to agricultural croplands or urban development?
2. How can water harvesting be used to augment local water supplies?
3. Since much of the arable land on earth has already been converted from wildland vegetation to croplands, why then do conversions from wildland vegetation to croplands remain an issue?
4. What is the role of BMPs in mitigating NPS pollution from a watershed and are they sufficient to address cumulative effects of a changing watershed landscape? Explain what a PMZ does in IWM?
5. What are the regulations in the United States that a watershed manager must consider in planning and implementing watershed management practices? Also, explain how compliance with these regulations can be achieved while meeting the goals of IWM?
6. What are the possible impacts of future climatic variability on hydrologic processes and the management of watershed?
7. Watershed managers are faced with the problem of having to make decisions when limited information and data are available. How can this situation be overcome?

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http://water.usgs.gov/nawqa/ U.S. Geological Survey National Stream Water Quality Assessment Program (accessed January 30, 2012)

CHAPTER 15 Socioeconomic Considerations in Integrated Watershed Management

INTRODUCTION

Earlier chapters of this book dealt with the biophysical aspects of integrated watershed management (IWM). To implement and sustain land-use changes or other actions needed to achieve the objectives of IWM, however, requires the consideration of socioeconomic factors that provide the incentives and ability to transform ideas and plans into actions. Proposed management practices are likely to fail if they do not recognize the wants of the people and the social situation in which these people live. Therefore, we consider in this chapter how socioeconomic factors and tools come into play in making sound decisions about IWM practices. In this chapter, we will refer to management *projects* as representing collective management practices that are often intended to achieve multiple benefits on a watershed. The following discussion would apply, however, to an individual practice as well as a project.

The reader is directed to Gregersen et al. (2007) for a more comprehensive discussion of the topics presented in this chapter that focus on

- how policies relating to watershed management evolve;
- how plans for implementing watershed management projects or individual practices are developed;
- reasons why an economic appraisal is often necessary to support a proposed watershed management practice; and
- methods of performing an economic appraisal that includes on-site land production, environmental consequences, and off-site effects.

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

Policies and Policy Processes

IWM is practiced in a policy context. There is always a framework of laws, regulations, institutional arrangements, cultural and social mores, and market conditions that govern and guide watershed management practices. We use the term *institutional* in its broadest sense to include all of the ways that people come together to cooperate, coordinate, and guide their activities. Within this context there are political, social, and economic challenges facing those attempting to improve the effectiveness and efficiency of watershed management practices. A further complication is the fact that policies are always changing and improving as more is learned. Furthermore, policies are often altered as the interests of additional people come to predominate and management priorities change. No matter what the status of a particular policy might be, however, its basic components should remain unchanged to be effective.

A Policy and Its Components

A policy is the establishment, implementation, and enforcement of the appropriate institutional arrangements to guide a land-use choice made by the people involved in making the choice. An effective policy specifies how to manage a project selected for implementation (Bromley, 1992). A policy consists of three essential components – intentions, rules, and compliance. Intentions reveal what the people hope to accomplish with the institutional arrangements confronted. Rules are the institutional arrangements that constrain some people and literate others. Compliance is the action that converts promises and proclamations into meaningful results.

The Policy Processes

People dealing with a proposed IWM project should understand the policy environment and the decision-making processes confronted before implementation can occur. They need to develop answers to a number of questions including

1. How do decision makers choose the issues to deal with and which policies to establish in response to issues?
2. How can the different people who are involved coordinate their actions when there are conflicts among groups of people and overlaps and interactions between decisions and policies?
3. Once decisions are reconciled into a set of policies, how can they be implemented while recognizing that policies represent decisions and not actions?
4. How can policies be enforced when voluntary action is required by most of the people affected?

The policy process is not a one-time process, but rather it is an interactive process of successive approximations as people move toward the long-range goals on which a consensus is reached. Policy changes are continuously taking place, often incrementally, with only a few major changes often defining new policy. Developing a policy for watershed management is largely similar to that in other sectors. Figure 15.1 provides an outline of the policy developmental process in general terms. Each stage of this process is discussed briefly below.

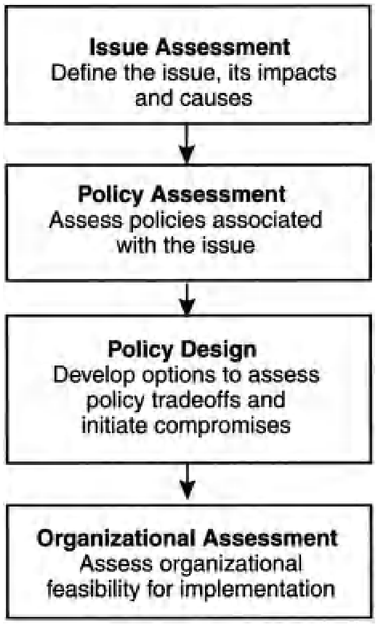

FIGURE 15.1. The process of developing a policy for watershed management

Identifying and Assessing Issues

There are many possible problems and opportunities related to a watershed management project that can warrant a policy response. Many of the issues confronted deal with policy objectives concerning

- achieving desirable land-use stability and soil conservation;
- reducing flood damages;
- sustaining water yield and minimum streamflows;
- maintaining or improving water quality;
- providing fair and equitable water allocations; and
- achieving sustainable economic growth and development.

Issues to be addressed by a policy arise when one or more of these objectives are not accomplished or when the conditions related to the objectives are worsening. Similarly, policies can be developed to respond to opportunities to improve people's welfare for people living both on the watershed itself and off the watershed but are affected by a watershed management project. Two policy issues that are related to IWM are

1. Watershed inhabitants are people with different perceptions about the best uses of watershed resources. Furthermore, the uses of these resources are likely to be valued differently by the people living on a watershed than those who live off the watershed, whether in uplands or downstream of the area to be impacted by a watershed project.

2. There can be disagreements and controversies among competing interests on the current and planned uses of watershed resources and the societal goals to be achieved by these uses.

Policy issues differ from technical management issues in terms of how they should be addressed. In the case of management issues, technical experts analyze the situation confronted and make recommendations for resolution of the issue. In the case of policy issues, there can be a disagreement over the management goals that lead to the emergence of the policy issue. Different sets of criteria for choice among options to resolve an issue also exist because different value frameworks exist. These differences need to be addressed as a first stage in the policy development process.

Policy issues also evolve through time as people's views on the uses of watershed resources change. For example, the conversion of large forests to agricultural production was not much of a policy issue in the early 1900s. However, this change in land use has become a major issue in later years as technologies have advanced, incomes have grown, comparative political power has shifted, and changes in the values and influence of environmental groups have taken place.

It is important to understand the different value perspectives and valuation frameworks of the people involved in watershed management and establishing a policy or policies. We need to identify and analyze these differing perspectives and the expectations of the different persons or groups of people involved in the issue. Here, the focus is placed on the following questions:

- From whose point of view is the identified issue a problem?
- What are the various opportunities and threats that the issue poses?
- Who benefits and who gains?

Once any disagreements have been resolved, a discussion and often a debate generally takes place in attempting to arrive at a policy structure that is acceptable by the people involved. Considerations of regulatory actions, market solutions, or combinations of both are part of these discussions. The takeoffs for all of the people involved should also be defined, debated, and negotiated.

Defining Solutions and Proposing Policies

Once general agreement on the policy issues is reached, a more detailed analysis takes place to define the takeoffs involved to accomplish the objectives. Then there is a need to understand the compensations needed to make the takeoffs acceptable to all people. Policies are needed that will fit the viewpoints of the people involved.

A result of the discussion leading to a policy might be the possible need to revise a policy or set of policies. However, the existing policies must be carefully assessed in terms of meeting their respective intentions before proposing the revised policy options. There can be two types of weakness in existing policies: (1) they are ineffective, that is, when a policy does not accomplish its objective, and (2) there are conflicting policies where a policy in one sector has unintended and unanticipated negative effects in another sector.

After analyzing the existing policies and finding a need for revision, the next task in the process is to identify the specific policies that require change. This task represents a transition to the third stage of the policy process.

Formalizing and Implementing Policies

Policy proposals generally move to an authoritative body or administrators of the implementing agencies if it is a matter of strengthening the regulations for enforcement of existing policies. At this point, there will likely be a discussion about whether *policy instruments* are needed for more effective enforcement of existing policies or whether the policies should be revised. These instruments fall into three categories: (1) regulatory mechanisms that include restrictions on land uses, land and resource rights-of-use, or water quotas or controls placed on damaging vegetation management practices; (2) fiscal and financial mechanisms such as higher taxes on excessive water use, subsidies for soil conservation practices or fines for improper management practices; and (3) direct public investment and management, such as increased or decreased levels of management of an infrastructure and resources, public conservation education, or specified research activities. Often, a combination of these three policy instruments will be most effective. This possibility is illustrated for the watershed-related problems in Box 15.1.

The implementation of either existing or revised policies depends largely on the nature of how the policies were established and the ways in which the people are involved in the implementation process. The main policies affecting watershed management in the USA are often established by government agencies. However, conflicts among the government agencies and between government agencies and the private sector can occur with one

Box 15.1

Watershed-Related Problems and Policy Actions

	Examples of alternative policy actions for specific problems		
Policy instruments	**Variable water supplies**	**Water shortages**	**Declining water quality**
Fiscal incentives	Use peak load pricing	Use opportunity cost pricing	Make polluter pay for damages
	Legalize water markets	Legalize water markets	Use tradable pollution permits
Regulatory mechanisms and institutions	Use floodplain zoning	Establish water user associations	Set water quality standards
	Establish priorities for water use in droughts	Establish river basin entities	Establish land-use zoning around streams and in watersheds
		Impose restrictions on use	
Direct public investments	Public groundwater development	Transfer water from surplus regions	Install waste treatment plants
	Expand reservoir storage capacity	Install water meters	Install aerators on polluted rivers
		Install water-saving toilets	

Source: Easter et al. (1995).

pursuing one approach and others pursuing other approaches. Therefore, consistency and coordination are essential.

Policies must be implemented and enforced to be effective. In some cases, institutional strengthening might be necessary to provide for effective implementation. It needs to be remembered that most of the ongoing watershed management practices in the USA are undertaken voluntary by landowners and resource users largely in response to incentives and management guidelines for the best management practices.

Monitoring and Evaluating Policy Implementation

Monitoring and evaluating policy implementation is an integral part of the policy process. People need to know how watershed conditions might be changing so that policies can be adjusted or otherwise modified to conform to the newer conditions confronted. Monitoring and evaluating of policies involve addressing three groups of issues relative to the situation and set of decision makers:

1. Who gains and who loses with respect to any decision and how to increase benefits and participation for any decisions is of interest. Included within this issue are impacts on different groupings of people, for example, people grouped by income classes, by regions or location, by gender, age, or occupation type.
2. Whether the gains in welfare are sustained over time must be determined. Included here is a combination of adequate resources to meet basic needs, security of access to these resources, and the maintenance of resource capabilities on a long-term basis.
3. Recognizing the concerns about possible changes in total benefits and costs because of the implementation of a watershed management practice or project must be addressed. These concerns are often measured by people's willingness to pay for the project and the relationship between incremental benefits and costs; that is, measured by *economic efficiency measures* to be discussed later in this chapter.

There can also be questions of concern about monitoring and evaluating the progress in a watershed management project within each of the issues listed above and, therefore, the effectiveness of policies that guide the watershed management project. Questions relevant to the first issue include

- How are the benefits and costs distributed among different groups of people?
- Who loses as a result of implementing any given watershed management practice and how are the losses occurring?
- In what way does watershed management encourage the equitable distribution of benefits and costs between the local and regional groups of people?
- How does watershed management affect local empowerment and participation in resource development?
- In cases where the success of a project is dependent on groups of people who are involved financially in the management of the project – is this involvement acceptable to them? What budgetary impacts will be confronted?
- How does watershed management impact other sectors and the stability in the immediate area and region?

Questions addressed with the second issue might be

- What are the impacts on the security of people's livelihood including their resource control and income generation?
- Are the benefits going to be sustainable through time?
- What are the budget and financial sustainability implications?

The third issue deals with the basic question – what is the relationship between the economic costs and benefits associated with the watershed management practices used to implement policies? This topic is discussed later in this chapter.

A Watershed Management Framework and Its Policy Implications

Institutional mechanisms are needed to help achieve the goals of IWM. Institutional mechanisms such as regulations, market and nonmarket incentives, public investments in research, education, and extension can ease the implementation of watershed management practices in many instances (Quinn et al., 1995). To be successful, however, these institutional mechanisms require a policy that recognizes the existence of the watershed and the processes functioning on a watershed. Furthermore, mechanisms are needed that allow for the accounting of the environmental benefits and costs associated with a watershed management practice.

Unfortunately, there are a several reasons why achieving successful watershed management through policy intervention is difficult, including

- Watershed boundaries and political boundaries rarely coincide. That is, biophysical processes and social, political, and economic activities do not normally take place on the same scale or within the same boundaries.
- Watershed management responsibilities are often shared by a multitude of organizations including some with management responsibilities of watershed resources, economic development, and extension activities. In practice, no single organization has coordination responsibility or authority.
- The full array of watershed management goals and benefits are often not totally considered, are not adequately understood, or given a low priority in formalizing the policies that deal with the management of natural resources. Consequently, these policies can hinder the practice of watershed management although this might not be intentional.
- Policies that encourage a land-use activity on a watershed are frequently formulated to increase the level of production of commodities with little consideration for the sustainability of the land-use activity, the desired level of production, or the negative off-site effects that can result.

Developing a policy to encourage IWM generally requires that landowners, government agencies, private companies, and other interested people become involved in making decisions. The general public must become aware of the importance of sustainable use of land, water, and other natural resources and the linkages upon which watershed management is built. All groups of people involved with the use of watershed resources and their perceptions and attitudes need to be identified. Who has jurisdiction over watershed management activities needs to be clarified as well as those with coordinating roles of the organizations involved in the management of watersheds? A key to developing appropriate policies favorable for IWM is that the policies promote the internalizing of off-site and downstream effects by more equitably distributing the benefits and costs associated with the use of the natural resources on watersheds. Therefore, the impacts of the policies related to IWM must be continuously assessed as they evolve through time.

Planning and Implementation

A management practice or project implemented on one area of a watershed often impacts on the land-use activities in other areas and, in doing so, can have positive and negative impacts on different groups of people. It is critical, therefore, that IWM practices being implemented are selected through an effective planning process. Planning helps to encourage and take advantage of the positive ones and avoid the possible negative impacts of others.

An Overview of Planning for IWM

Planning for IWM involves three sets of factors: (1) objectives must be established on the basis of a problem analysis; (2) constraints to implementing a proposed management practice (project) need to be determined, including biophysical limitations, budget restrictions, and social, cultural, political conditions that are associated with the situation; and (3) there needs to be managerial techniques and capabilities available for implementing any proposed practice. Planning, therefore, involves the integration of objectives, constraints, and available techniques to improve the effectiveness and efficiency of implementing IWM practices. The art of integrating biological, physical, and social sciences is emphasized here since planning is as much art as a science. Each watershed-related issue has its own unique set of technical characteristics and each management practice can require different technical approaches. The same does not hold for the process of planning. A similar planning process can be used regardless of the type or number of watershed management practices being proposed. It is only the emphasis placed on each step in the process that will differ.

A basic principle in assessing both the positive and negative impacts of an IWM project is application of what is called the *with-and-without concept*. In other words, we want to assess the changes that occur with and without any given watershed management project. For example, when we talk about reducing sedimentation in a downstream reservoir, we are referring to the difference in sedimentation with and without erosion-control practices. Sedimentation can still be occurring with a given erosion-control practice but at a slower rate than without the practice. Not to be neglected, however, are the effects of the project on upland productivity where soil losses are avoided (and productivity sustained over that without the project). Differences with and without the project are illustrated in Figure 15.2.

A related principle is that losses prevented by a watershed management practice have to be treated in a similar way as gains obtained when applying the with-and-without concept. For example, sedimentation in the reservoir can still be increasing following implementation of the management project but at a slower rate than without the project. This is an important principle because so many of the benefits of IWM are losses prevented rather than net gains.

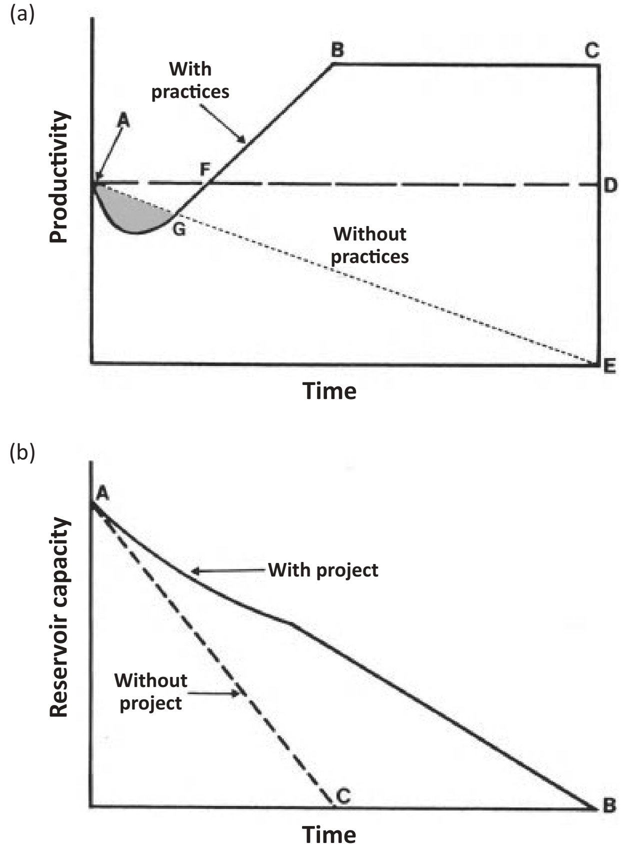

FIGURE 15.2. The relationship of upland productivity (**a**) and sedimentation in a downstream reservoir (**b**) under conditions "with" and "without" watershed management practices (from Gregersen et al., 1987; © FAO, with permission)

Steps in the Planning Process

The planning process for watershed management is likely to involve the following five sequential steps:

1. Monitor and evaluate past activities and identify problems and opportunities.
2. Identify main characteristics of the problems confronted, opportunities to resolve these problems, and establish the objectives and constraints to accomplish these objectives. This then leads to the formulating strategies for action.
3. Identify alternative management practices (projects) to implement the formulated strategies within the limits of the constraints.
4. Appraise and evaluate the impacts of the alternative management practices (projects) including the environmental, social, and economic effects and assess the uncertainty associated with the impacts.
5. Rank or otherwise prioritize the alternative management practices and recommend the project to be implemented when a recommendation is requested.

Evaluating Past Activities and Identifying Problems and Opportunities

The planning process has no beginning and no certain end. A logical starting point, however, is before a problem is identified through monitoring consisting of careful observation and measurements and evaluations of resource responses to climate, management, or the lack of management. Often, no formal monitoring or evaluation system is used to obtain the information that leads to identification of a watershed issue of concern and eventual action. Instead, problems are often observed only after they have occurred such as when the scars of erosion begin to appear on a landscape, when a reservoir is silting up rapidly, or when flooding becomes more frequent. Opportunities to mitigate the problems are then identified.

Regardless of how problems and opportunities are recognized, their articulation becomes one of the first steps in planning. In many instances, more than one solution to a problem is possible; for example, insufficient water supplies for downstream users might be enhanced by increasing the flows of water from an upstream watershed or developing reservoirs downstream to store streamflow for future use. In other cases, some solutions are mutually exclusive with one action precluding another. While the specific actions taken in each case might differ, the planning process remains largely the same.

Establishing Objectives, Identifying Constraints, Developing Strategies

The next stage in the planning process involves establishing objectives and identifying constraints in developing strategies to solve the problems or to take advantage of the opportunities. Objectives generally evolve from the problem analysis. Statements of objectives indicate that there is a need to develop an effective response for overcoming or preventing the problem. A single objective or set of multiple objectives (Box 15.2) are then translated into actions that can be constrained by the riskiness of approaches taken, the level of cost necessary, and the level of success in accomplishing other objectives.

Once objectives have been established and constraints identified, a general strategy for action needs to be developed. The important thing here is not necessarily the strategy statement itself but the process by which it was developed. If we only look at the problem statement, we likely could think of a number of alternatives to solve the problem confronted. For example, we could suggest a watershed management practice involving the conversion of one forest type to another to increase streamflows. In other cases, the best strategy might be to leave the situation alone and spending our resources elsewhere, such as developing groundwater resources or diverting water from a water-rich to a water-poor area.

Identifying Alternative Ways to Implement the Strategy

After an acceptable strategy has been developed, we get down to the details of evaluating the alternative projects that could be implemented. The need here is to identify the possible watershed management practice or project that could be used to successfully implement the

Box 15.2

A Single Objective versus Multiple Objectives
Even when constrained in many ways, a single objective is relatively easy to consider within the planning process. Clearly stated decision criteria can be developed for determining the extent to which alternative watershed management practices might be acceptable and how these management practices rank relative to each other. Difficulties arise when more than one objective exists. Decision-making models have been developed to deal with multiple objectives (El-Swaify and Yakowitz, 1998, and others). Some planners avoid the problem of dealing quantitatively with multiple objectives by developing an array of alternative management practices to accomplish the objectives. The planner then has to compare the relative merits of the alternative practices to make a decision on the best action.

strategy to obtain the results desired. This is the place in the planning process where technical specialists, social scientists, decision makers, and others dealing with socioeconomic and cultural issues come into the picture. The task of the planner is to identify the possibilities and the array of options that are available within the constraints and circumstances surrounding the management project.

Appraising Alternatives

While alternatives are being developed, they are also being appraised. In its broadest meaning, the term *appraisal* refers to the process of identifying, defining, and quantifying the likely impacts of the watershed management practices. Some of the possible effects of these practices are listed in Table 15.1. The separation of these impacts into economic and financial, environmental, and social effects relate to the different types of effects that a change in watershed management can cause.

Making Appraisals Useful

Appraisals of proposed watershed management practices and projects are useful only if they provide timely information of relevance to the planner, manager, and decision maker. A distinction often needs to be made between the technical analyst's considerations in choosing an appraisal procedure and the people's viewpoint of what characterizes an acceptable appraisal of the alternatives. A task of the planner is to bring these two perspectives together in the final appraisals.

Appraisals of watershed management practices are pursued generally in sequential levels of analysis because the resources available for these appraisals are limited in most cases. Starting with only two alternatives, for example, one management action and the option to do nothing, is generally too restrictive. The preferable approach is to start with a number of alternative management practices and then to narrow them down systematically

Table 15.1. Scope of effects of watershed management practices

Economic/financial effects on
- Regional and national level of production
- Allocation of resources
- Regional and national income
- National balance of payment
- Stability of income over time
- Distribution of income (both interpersonal and intertemporal)
- Public budgets

Environmental effects on
- Ecological diversity
- Ecological stability
- Wildlife protection
- Soil protection
- Landscape aesthetics
- Water yield and timing
- Water quality
- National patrimony

Social effects on
- Regional employment
- Working conditions
- Public participation
- Migration flows
- Cultural traditions
- National vulnerability
- Political stability

Source: OECD (1986), with permission.
Note: For convenience in exposition, we have divided the effects into three categories, namely, economic/financial, environmental, and social effects. Other categories could equally as well have been chosen.

in stages. This approach also encourages the introduction of economics into the planning process rather than tacking it on at the end of the planning process through a feasibility study.

Risk and Uncertainty of Appraisals

A planner faces a situation of uncertainty rather than risk with appraisals of most watershed management practices. One can apply probabilities to various outcomes in the case of *risk*, while measures of the probability of occurrence cannot be generated in the case of *uncertainty*. One can also develop *subjective probability estimates* for different aspects of the management practices in a situation of uncertainty. However, such estimates might do more harm than good since subjectivity in the planning process should not be hidden. We suggest using a *sensitivity analysis*, which is an analysis of how the measures of worth (value) or desirability of the alternative management practices change under different assumptions concerning the values of key parameters of the practices to be appraised (Gregersen et al., 1987).

Recommending Action

In some instances, a planner's task stops when the alternatives and the implications of risk and uncertainty for the alternatives have been evaluated. In other cases, however, the planner might be asked for recommendations on which of the alternative management practices should be selected for implementation and the timing and approach to its implementation. To this end, the appraisal results can be presented to the decision makers in different ways depending on the planning situation. Offering a ranked set of alternatives is often preferable or perhaps several rankings utilizing different appraisal criteria should be presented. Importantly, only the responsible decision maker can decide ultimately which alternative is chosen.

Planning as a Continuous Process

Planning is a continuous process with information concerning results of the management actions taken and emerging problems continuously fed back into the process. This information is then used to suggest possible changes in the ongoing watershed management practice. The process of collecting and disseminating information relating to ongoing management practices is part of the monitoring and evaluation effort discussed earlier. This continuous process leads to valuable interactions among planners, technical personnel, and managers of watersheds and relevant decision makers.

Economic Appraisals

While watershed managers often deal mostly with biophysical processes, they should also have an understanding of the economic and financial implications of what they are doing. It is important to know the economic values of each of the alternative watershed management practices considered to be feasible for implementation in the planning process, how much each of them will cost to implement and maintain, and the cash flow required to sustain each of the management practices through time. These people will also want to know who is responsible for obtaining and allocating the financial resources necessary for the practices. Therefore, we explore the economic aspects of watershed management in this section of the chapter. An overview of the ways in which economists value and analyze the benefits and costs associated with watershed management practices is also presented.

Distinction between Economic and Financial Efficiencies

One point that should be clarified at the beginning of this discussion is the distinction that is made between economic and financial efficiencies. *Economic efficiency* relates to the relationships between total benefits and costs that are valued in either market prices or economic values when market prices are not appropriate. *Financial efficiency* is concerned with *profit*, that is, the difference between market-priced benefits and costs. Therefore, a financial analysis deals with market-traded goods and service, money inflows and outflows, and who gets paid and who pays. An economic analysis also considers these issues, but it attempts to value them in terms of society's willingness to pay for them. An economic analysis includes social benefits and costs of services not traded in the marketplace such as flood prevention and the maintenance of healthy riparian communities. We will focus on economic efficiency because it better reflects people's interests in watershed management that are usually beyond only income generation.

Time Value of Money, Discount Rate, Present Value

Economists often look at economic efficiency in terms of present values; that is, all of the estimated future benefits and costs are brought back to a common point in time that most generally is the present. This task is accomplished by discounting the benefits and costs to a common point by applying a *discount rate*. If the present value of benefits is greater than the present value of costs for a watershed management practice and there is no cheaper way of achieving the objective being sought, the management practice is considered to be an economically efficient use of the available resources.

Discounting assumes that society values a dollar of present benefit or cost more highly than a dollar of benefit obtained or cost paid in the future. When we lend money as a business, we expect to get interest as a payment for forgoing the use of the loaned money today. For example, if we lend \$100 at 8% interest, we expect in 10 years to get \$100 $\times$ $(1.08)^{10}$, or \$215; this is our original \$100 plus \$115 of interest. We could get this value using the simple compound-interest formula. Therefore, future value is determined by

$$V_{\mathrm{f}} = V_{\mathrm{p}}(1 + r)^t \qquad \textbf{(15.1)}$$

where V_{f} is the future value; V_{p} is the present value; r is the annual compounding rate, or interest rate; and t is the time from year 1 through year n.

In contrast, we could discount \$215, for 10 years at 8% interest by the discount formula and arrive back to the \$100 of present value. Present value is determined by

$$V_{\mathrm{p}} = \frac{V_{\mathrm{f}}}{(1 + r)^t} \qquad \textbf{(15.2)}$$

For the above example, $\$215/(1.08)^{10} = \$100$.

Steps in the Economic Appraisal Process

An economic appraisal of a proposed watershed management practice or project is undertaken to assist decision makers to determine which of the watershed management practices deemed to be feasible in the planning process should be implemented. If only one management practice has been selected in the planning process, an economic appraisal would help to indicate its economic value. The economic appraisal process for watershed management includes the following four steps (Gregersen et al., 1987) for each alternative identified in the planning process:

1. Identify and quantify the physical inputs and outputs involved and determine how they occur over time.
2. Determine values of inputs and outputs and the changes in these values over time.
3. Compare the valued costs and benefits by calculating relevant measures of project worth.
4. Consider the nonmonetary (e.g., environmental services) benefits and costs when necessary.

The framework in Figure 15.3 is useful for depicting the physical effects of a watershed management project that can include multiple practices to achieve increases in outputs such

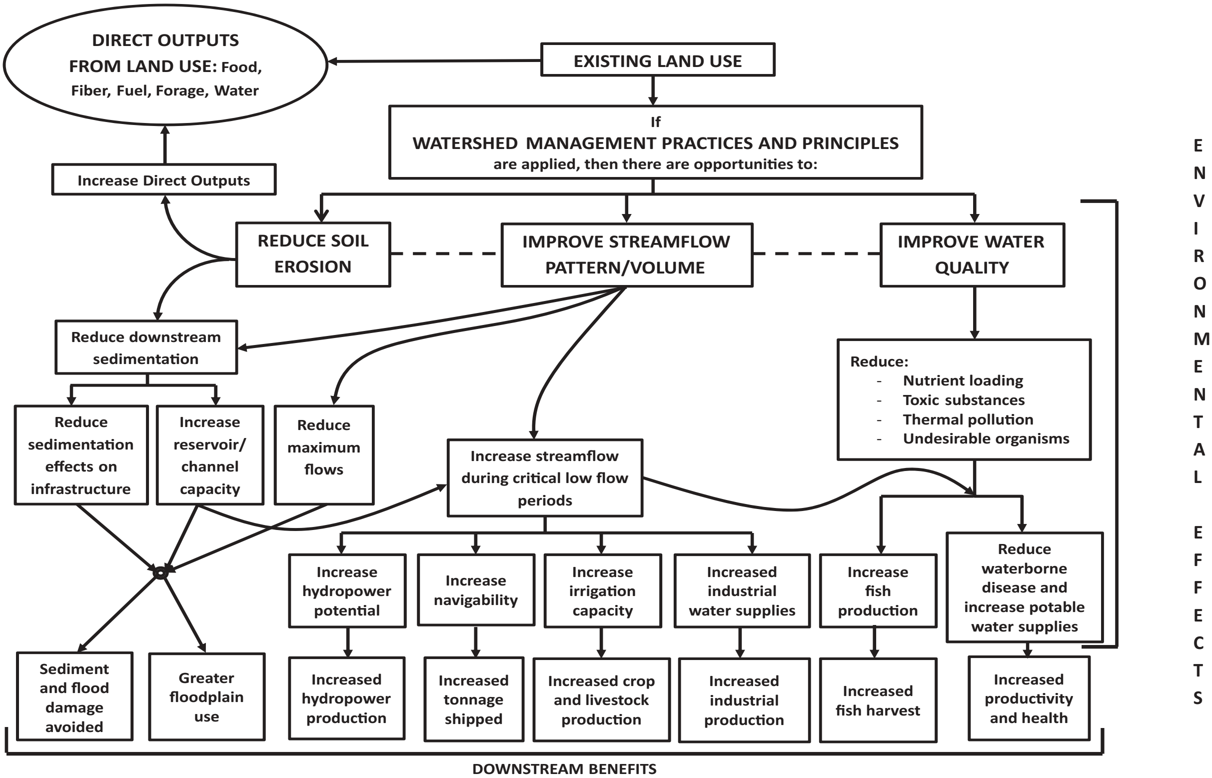

FIGURE 15.3. Framework for examining relationships among physical effects, environmental changes and downstream benefits "with" watershed management practices as compared to conditions "without" the practices (adapted from Gregersen et al., 1987)

as production (food, fiber, forage), environmental protection, and enhancement of water resources. It also serves as a guide for quantifying and valuing inputs and outputs that occur on the watershed and downstream. Using this framework, a planner can identify and trace the effects of a project on production, soil stability, sediment flows, water yields and pattern of water flow, and water quality. Such a framework helps identify the possible effects of a given project that can differ for the physical–biological and socioeconomic setting in which the project occurs; therefore, all the items indicated in this figure would not necessarily apply for any given assessment. This framework allows the planner or manager to identify key outcomes that are intended and translate the outcomes into terms (value measures) that are meaningful to economists in making the economic appraisal. The framework is applied for conditions with and without the project being implemented to provide values of assessing the economic worth of the project.

The economic appraisal process is generally iterative with the analyst passing through increasingly detailed stages of evaluation. Each of the sequential steps in an economic appraisal is discussed in the following paragraphs.

Identifying and Quantifying Physical Inputs and Outputs

Obtaining information on the physical inputs and outputs of a particular watershed management practice or project and the relationships between inputs and outputs is one of the major tasks in the appraisal process. It is in this first step that the primary interaction between a technical watershed management specialist and an economic analyst takes place. The technical watershed manager has to identify and quantify most of the inputs and associated outputs in physical terms and defining the watershed management alternative(s) to be evaluated. For an economic analysis, this information is needed in terms of

- the physical units in which the inputs and outputs are measured;
- the source of the inputs; and
- when the inputs will be needed and when the outputs will occur.

Watershed managers can generally meet these informational needs if they are aware of them early in the planning process. Examples of inputs required in implementing a watershed management project are shown in Table 15.2. The inputs must then be related to the outputs from which they are derived. This information can be summarized in a *flow table* that shows the flows of the physical inputs and outputs through time.

A point that bears emphasis is that the input and output quantities in the physical-flow table should reflect the differences with and without the management practice (see Fig. 15.2). Many outputs obtained with through IWM practices are expressed in the form of *losses prevented*, for example, in the losses prevented through flood prevention, or preventing degradation of riparian vegetation. Even though these outputs can be difficult to quantify in applying the with-and-without principle, the relevant outputs should be included in the appraisal process because they are important in *human-value* terms. As a consequence, the tons of soil loss prevented by implementing an erosion-control practice are not an adequate output measure. People normally do not put a value on the soil loss. Therefore, the soil loss prevented must be related to a human value, for example, to food production or other losses prevented that can be linked to human values.

TABLE 15.2. Examples of inputs for a watershed management practice or project

Category of inputs	Examples/description
Workforce	Resource managers – forest, range, watershed managers and planners Engineers and hydrologists – design of erosion-control structures, floodplain analyses, water yield estimate, etc. Skilled labor – construction Unskilled labor Training/extension specialists to facilitate adoption of project
Equipment	Detailed listing of equipment needed for project construction and maintenance Schedule of needs and equipment maintenance, i.e., timing
Land	Land classified according to suitability for various uses Designate sensitive areas to be protected (benefits foregone) Areas to receive treatments followed by management
Raw materials	Utilities (energy, fuels, etc.) Wood (construction, fence posts, etc.) Other construction materials (concrete) Water
Structures and civil works	Housing, roads, other facilities needed for project that are not part of project itself; if part of project, they are included in work force, materials, etc., listed above

Source: Gregersen et al. (2007); © FAO, with permission.

The values of some of the outputs of watershed management such as preventing water quality impairment or aesthetic benefits are not easily measured in the marketplace. Even if these outputs cannot be quantified and valued, however, they should be described to the extent possible in an economic appraisal. Many nonquantifiable benefits such as ecosystem preservation or the protection of biodiversity relate to the issues associated with the sustainability of human activity. Benefits such as these are also important but which cannot be quantified and valued other than in a descriptive fashion.

A discussion of input and output relationships is complete when the relationships between inputs used and goods and services consumed and valued by society have been established.

Valuing Inputs and Outputs

Prices established at the marketplaces are used to value the inputs and outputs of watershed management practices that are traded in the market. We, therefore, will concentrate on the valuation of nonmarketable benefits and costs through shadow pricing.

Measures of Economic Value and Shadow Pricing. A basic measure of economic value is the *willingness to pay* (WTP). WTP is a measure that reflects society's willingness to pay for goods and services at the margin, that is, if another unit of the good or service was made available. The WTP is a reflection of *scarcity value* in that the more that is made available, the less any individual generally is willing to pay for the good or service at the margin.

Another common measure of economic value used is the *opportunity cost* (OC), which is a measure of value of the opportunity foregone when a resource is used for one purpose rather than another. An example would be the OC of setting aside a parcel land as a watershed protection area. The OC applied to the land set aside for this purpose would be the value of timber, livestock production, or other benefits foregone. There is a relationship between the WTP and the OC. OC values are used in measuring the WTP for the goods and services foregone.

One can assume that market prices reflect the WTP at the margin in a competitive economy with no constraints on the movement of prices. This is the reason why market prices are widely used in economic analyses. However, the WTP and OC can diverge from market prices when regulations are placed on the market prices such as when subsidies or taxes affect them (Gregersen and Contreras, 1992). When this divergence occurs, market prices are adjusted to reflect true scarcity in the economy. These adjusted prices are called *shadow prices*. An example of where shadow pricing is required is when the goods and services do not have observable market prices. Many environmental services obtained in a watershed management practice are of this type. The economic analyst attempts to derive shadow prices that reflect society's WTP for the good or service in this case.

Developing Shadow Prices. One of three approaches is used in developing shadow prices for the benefits and costs of a watershed management practice: (1) market prices, (2) surrogate market prices, or (3) hypothetical valuation approaches.

Where a market price is considered to adequately reflect the WTP for a good or service, the market price itself can be adjusted to reflect the value of the good or service at the margin. Reasons for this suggestion include

- Market prices are accepted more readily by decision makers than artificial values derived by the analyst.
- Market prices are generally easy to observe at a point in time and through time.
- Market prices reflect the decisions of many people acting as buyers and not only the judgment of the analyst that is the case with subsidized prices.
- The procedures for calculating shadow prices are imperfect, and, therefore, estimates of shadow prices can introduce larger discrepancies than the simple use of even imperfect market prices.

Surrogate market prices are used in situations where benefits and costs are not themselves valued in the market but for which clear substitutes exist in the marketplace. The market prices of the substitutes are often used to develop surrogate values for the benefits or costs to be valued. For example, there is no market for the amount of soil that is eroded from an upstream watershed or the amount of sediment that is deposited in a downstream reservoir. However, one approach to placing a value on the eroded soil or the deposited sediment is to examine the market prices of the eroded uplands or the silted lowlands and then compare these prices to the market prices of comparable lands that are not affected by erosion or sedimentation. The difference in respective land values acts as a surrogate price for the damage caused by the erosion or sedimentation.

Hypothetical valuation approaches apply when it is not possible to derive acceptable market price measures of value. In such cases an economist might derive some information about value through local surveys or expert judgment to estimate the minimum values for some of the benefits of a watershed management practice through an analysis of the cost of obtaining these benefits. This approach is a *cost-price analysis* since it uses costs to derive some information to estimate the minimum value of the benefits that would be required to break even with costs. It must be kept in mind, however, that such values represent a lower value of the good or service in question. In other words, the resulting value is the minimum value the decision maker would have to place on the output to ensure a break-even relationship between economic costs and benefits.

Because neither surrogate nor hypothetical market prices rely on actual market prices, the results obtained in applying these approaches to place a value on goods and services must be interpreted carefully. These approaches should be used in an economic analysis only when market-based approaches are impractical or impossible to apply.

Further details on these valuation methods are found in Gregersen et al. (1987), Winpenny (1991), Dixon et al. (1994), and others.

Comparing Costs and Benefits: Measures of Project Worth

Once the physical-flow tables and the associated values are formulated, the two are brought together to develop an *economic value-flow table*. An economic value-flow table lists the flows of economic benefits and costs through the expected life of a watershed management project. A *cash-flow table* would also be developed if a financial analysis is undertaken. The implications of risk and uncertainty must be considered in these calculations, for example, by a *sensitivity analysis* that indicates how the measures of worth might change with changes in assumptions concerning the input and output values.

The reason for the development of an economic value flow is to organize information to compare and evaluate the feasibility of the project alternatives. Two evaluation questions are always of interest to decision makers: (1) is the proposed watershed management project worth doing, and (2) if so, is the project better than other alternative uses of the available and often scarce resources?

Additional questions concerning benefits and costs that are relevant in the appraisal of a watershed management project are

- What is the budget impact likely to be?
- Is the project that is selected for implementation attractive to all of the people who will have to put resources into the various practices to make them work?
- What are the income-distribution impacts of the proposed project?
- Once implemented will the project have balance-of-payment impacts?
- Are the economic benefits greater than costs? That is, is the project selected for implementation an economically efficient use of resources?
- Will the project increase economic stability of the affected region?

Answers to these questions help to place an economic appraisal of alternative watershed management projects into a proper perspective for analysis.

Once the economic value-flow table has been developed, an economic-efficiency analysis is undertaken to compare the streams of benefits and costs for the alternative watershed management projects selected in the planning processes. There are three value measures that can be used in an analysis of economic efficiency – the *net present worth*, the *economic rate of return*, and a *benefit–cost ratio*. All of these measures are calculated with the same benefit and cost data and assumptions.

Net Present Worth. Net present worth (NPW), also known as net present value, is based on a need to determine the present value of net benefits from a watershed management project. The use of the NPW criterion will provide a ranking of alternatives if the goal is to determine the total net benefits of a project or practice to society. The formula for the NPW calculation is

$$\text{NPW} = \sum_{t=1}^{n} \left[\frac{B_t - C_t}{(1+r)^t} \right] \tag{15.3}$$

where B_t and C_t are the benefits and costs in year t; and r is the selected discount rate.

Economic Rate of Return. The economic rate of return (ERR) is also used to evaluate alternative watershed management practices. Unlike the NPW or a benefit–cost ratio, the ERR does not use a predetermined discount rate in its calculation. Rather, the ERR is the discount rate that sets the present value of benefits equal to the present value of costs. The ERR is the discount rate, r, such that

$$\sum_{t=1}^{n} \left[\frac{B_t}{(1+r)^t} \right] = \sum_{t=1}^{n} \left[\frac{C_t}{(1+r)^t} \right] \tag{15.4}$$

or

$$\sum_{t=1}^{n} \left[\frac{B_t - C_t}{(1+r)^t} \right] = 0 \tag{15.5}$$

Although a discount rate is not prescribed but is determined as a result of the ERR calculation, this does not eliminate the use of a reference discount rate in its analysis. The calculated ERR is compared to a predetermined discount rate to decide whether a project is economically efficient. For example, if the ERR calculated is 15% and the OC of the funds necessary to implement the project represent 10%, the project would be economically attractive. However, if the funds for the project cost 18%, the project would be financially unattractive.

A Benefit–Cost Ratio. A benefit–cost (B/C) ratio simply compares the present value of benefits to the present value of costs:

$$\text{B/C ratio} = \frac{\sum_{t=1}^{n} \left[\frac{B_t}{(1+r)^t} \right]}{\sum_{t=1}^{n} \left[\frac{C_t}{(1+r)^t} \right]} \tag{15.6}$$

If the B/C ratio is greater than 1, the present value of benefits is greater than the present value of costs and the management practice is an economically efficient use of financial resources assuming that there is no lower-cost means for achieving the same benefits.

Deciding Which Value Measure to Use

Maximum total NPW is the economic objective we seek for the investment of available scarce resources. Therefore, the NPW measure should always be a part of any ranking scheme for accepting or rejecting watershed management practices, projects, or programs

(Dixon and Hufschmidt, 1986). The ERR and the B/C ratio are measures of benefits per unit of cost. However, these two value measures give no indication of the total magnitude of the net benefits. Since net benefit is what we want to maximize for a given investment budget, reliance on just the ERR or B/C ratio could lead to the selection of a project that provides total net benefits that are smaller than those resulting when projects are selected using the NPW criterion.

In cases where projects are not mutually exclusive and there are no constraints on costs, all projects that result in a positive NPW can be accepted. In cases where not all projects can be selected for implementation because of a cost constraint, the goal is to select that project or those practices that provide the greatest total NPW. For mutually exclusive projects, for example, when two or more practices that would use the same site, the NPW measure is the only value measure that will always lead to the correct selection.

Nonmonetary Benefits and Costs

In spite of all the advances that have been made in the economic valuation of nonmarketed goods and services, there are frequently some of the effects of a watershed management practice that are impossible to either quantify or value. For example, the construction of a capital infrastructure such as a dam can have a negative aesthetic impact to people in that the view is not as natural or pleasing as it previously was. This kind of aesthetic effect is almost impossible to quantify largely because there are no widely accepted measurement or expressions of value for scenic beauty. A practice or project that requires a change in the lifestyle of the local people would have a *cultural* impact. The impacts of these nonmonetary benefits and costs are also difficult to quantify in this example. Nevertheless, these nonmonetary effects should still be recognized and described within the analysis, evaluated qualitatively, and presented to the decision maker. These effects will not be ignored by the decision maker even though they cannot be entered directly into the economic analysis in this way.

Application of the Economic Appraisal Process

Applications of the process outlined by Gregersen et al. (1987) for the economic appraisal of watershed management projects have been presented by Brooks et al. (1982), Wang et al. (1998), and Shuhuai et al. (2001). The application by Shuhuai et al. (2001) is described here to illustrate the process.

Shuhuai et al. (2001) estimated the benefits and costs of a proposed watershed management project to protect a principal source of water for Beijing, the capital of the People's Republic of China, to determine if the project represented an economically efficient use of the resources available for the project. More specifically, the Beijing Water Conservancy wanted to learn more about the economic feasibility of assembledges of proposed watershed management activities for enhancing the livelihood of people living above the Miyun Reservoir, about 85 km northeast of Beijing, while maintaining a flow of high-quality water into the reservoir.

Proposed Watershed Management Activities

The proposed watershed management activities to possibly be implemented on an area of 3300 ha above the Miyun Reservoir included conservation measures on forests and farmlands, protection of the water resources in the reservoir, and proper construction of roads in the project area. Erosion control was emphasized with varying combinations of planting vegetative covers, improving farming practices, and constructing engineering structures. The specific components of the activities were

1. Tree cutting and livestock grazing were to be excluded on 670 ha of forests and nonproductive wastelands with slopes greater than 25% and soil depths less than 30 cm.
2. On lowlands with gullies, 1500 m of check dams and 10,000 m of "protection" dams would be constructed to
   - Reduce soil erosion on the hillslopes and, therefore, the sediment delivery to streams.
   - Attenuate flood peaks impacting downstream villages and farmlands.
   - Provide water to facilitate agricultural irrigation.
3. About 15 km of channels to drain floodwaters and 40 km of roads to improve transportation in the area would be constructed.
4. Tillage practices were to be improved to reduce soil erosion and increase crop yields on 300 ha of farmlands with slopes less than 10%.
5. Forests having economic value would be established on 410 ha of farmlands with slopes of 10–25% and on nonproductive wastelands with similar slopes.
6. Fruit trees would be planted on 590 ha of land considered "marginal" in terms of annual crop productivity potential.
7. Conservation forests were to be planted on 133 ha using fish-scale microcatchments to promote establishment where slopes were greater than 25%; the purpose of these forests was to stabilize hillslopes and thereby control soil mass erosion.

Inputs and Costs

The inputs and costs for the proposed watershed management activities were the labor to implement and management the practices, the materials required for the activities, and the opportunity cost of resources previously available to the local people for other uses. These data were made available by the Beijing Water Conservancy. These inputs were valued by either market prices or estimates of opportunity costs. Because the opportunity cost for labor was less than the wages for labor in the region, a sensitivity analysis was conducted to determine whether higher labor costs would affect the outcome of the analysis.

Outputs and Benefits

Expected outputs from the proposed management activities included the interlocking benefits of reduced soil erosion, environmental enhancements, and eventually increased levels of sustainable agricultural production. The anticipated on-site and off-site (downstream) benefits in comparison to estimated benefits obtained without the project included

- Increased agricultural crop production and a reduction in fertilizer costs were predicted through soil stabilization and improved soil conservation. Less fertilizer will be needed to maintain or increase crop productivity with the improved soil conservation, which will increase income for farmers.
- Costs of removing sediment in stream channels, ponds, and the Miyun Reservoir would be avoided. The anticipated reductions in on-site soil loss and consequent

delivery of sediment into stream channels are the reasons for this cost avoidance, which represent a benefit of the proposed project.

- Water quality was to be enhanced with improved drinking water for local people and people living downstream. The expenditures for water treatment and health maintenance will be reduced by improving the people's health and well-being.
- Flood protection and the reduction in losses of houses, farm buildings, and farmlands, and (in extreme cases) people's lives were considered to result from the project. People's income and well-being should be improved and, as a result, government expenditures should be reduced further.

Valuing Benefits

While estimates of the costs of the project could be made with relative confidence, knowledge of the economic values associated with the benefits was incomplete, imperfect, or nonexistent. The approach taken by Shuhuai et al. (2001) in dealing with the more difficult-to-measure benefits was to begin their analysis with the values for which "acceptable estimates" of benefits could be made. If the resulting comparison of costs and benefits indicated a positive value, the assumption was made that any additional benefits would only enhance the project and, as a result, the project should be acceptable from society's point of view. However, if the initial comparison is unfavorable, it would be necessary to reexamine the benefits to obtain "acceptable" quantitative estimates for the benefits that are more difficult to measure. Values were calculated (in monetary terms) for the following benefits:

- increased agricultural production from improved tillage and the establishment of economic and conservation forests;
- reductions in soil loss and sedimentation; and
- flood mitigation effects.

The benefit of the anticipated reduction in surface runoff and, as a consequence, the reduction in sediment delivery to stream channels as a result of the improved vegetative-cover conditions and increased terracing would have little effect on the 4.2 billion m^3 of storage in the Miyun Reservoir. Therefore, it was not included in the analysis. Because of the difficulty in quantifying the improvement in water-quality downstream, this benefit was also excluded from the analysis.

Comparing Costs and Benefits

It was determined that the proposed watershed management project would provide a 25% rate of return on the capital investment to be made in implementing the project. It was noted, however, that based on the comparisons between the estimated costs and estimated benefits for the duration of the project, a positive economic value would not be obtained until year 7 of the 20-year duration of the project. But the flow of positive economic values increases rapidly from that point onward. This result is characteristic of the long-term nature of most watershed management projects and emphasizes the need to take a long-term perspective in appraising the economic value of these projects. Conversely, projects that aim to show "large impacts" quickly to attract people's support might not lead to sustainable economic development.

A sensitivity analysis was included in the appraisal to determine the implications of uncertainty and, as a consequence, the validity of the assumptions made. Specifically, sensitivity of the calculated rate of return to changes in the project's costs and benefits was tested by

- assuming labor costs to be twice and five-times the values used in the initial analysis; and
- conducting additional analyses with the benefits being 10% and 25% lower than those in the initial analysis.

The sensitivity analysis indicated that the rate of return would remain above the 10% alternative rate of return if labor costs are doubled but there is a negative NPW when labor costs were increased by a factor of five. The rate of return also remained above the 10% alternative rate of return or the discount rate assumed to be relevant to this project when the labor costs were doubled or when the benefits are decreased by 10%. There were small differences between the two reductions in benefits attributed largely to the fact that the main benefits would occur later in the life of the project.

Results

The result of the appraisal indicated that unless labor costs increase substantially, the proposed watershed management project should merit public support. This conclusion is strengthened by the fact that some of the anticipated benefits of the project such as flood control and water quality are difficult to quantify and, therefore, were not considered in the analysis. By bringing these benefits into the analysis, the flow of benefits would be even more favorable to the project.

SUMMARY AND LEARNING POINTS

This chapter has dealt with watershed management policy and planning issues and what economists do to analyze the costs and benefits associated with a watershed management practice or project. This information is needed to consider the tradeoffs involved between alternative watershed management practices and between the best allocations of available resources. When you are finished with the chapter, you should understand

1. That an appropriate policy environment is a prerequisite for implementing a watershed management practice or project that satisfies the objectives specified by the interested parties.
2. The general stages in the policy development process and factors that influence IWM policy interventions.
3. The importance of planning watershed management practices and their coordination in achieving land-use changes on watersheds that have interrelated impacts both spatially and temporally.
4. What is involved in planning for watershed management? What are the steps in the planning process and how does one appraise watershed management plans and projects?

5. The nature of an economic analysis and the economic and social impacts of a watershed management practice or project.
6. The ways in which an economic appraisal is applied to evaluate alternative watershed management practices or projects.

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CHAPTER 16 *Tools and Emerging Technologies*

> Mis-use of these advanced technologies can be hazardous to your professional career.
>
> *Unknown Author*

INTRODUCTION

Many of the field and analytical tools currently used by hydrologists and watershed managers in improving their understanding of hydrologic processes and watershed values are reviewed in this concluding chapter. Emerging technologies that help in expanding this knowledge in the future are also considered in relation to their potential contributions to the scientific and management communities. These emerging technologies offer the promise of making research and management tasks easier and providing capabilities previously unknown.

Computer-based hydrologic simulation models, remote-sensing technologies, geographic information systems, global positioning systems, decision-support systems, internet applications, and isotope technologies are among the tools available for hydrologists and watershed managers. While these tools are described in this chapter, we will concentrate more on the development, capabilities, and applications of hydrologic simulation models because of their widespread and continuing use in hydrology and integrated watershed management (IWM). Furthermore, hydrologic simulation models are rapidly being improved with advances in computer technology that facilitates their capability to interface with the other aforementioned technologies, providing more powerful tools for operational and research applications. While such technological advances are impressive, we emphasize that the users of models should understand the capabilities and limitations of models in

Hydrology and the Management of Watersheds, Fourth Edition. Kenneth N. Brooks, Peter F. Ffolliott and Joseph A. Magner.

simulating and predicting hydrologic processes and responses to varying watershed and climatic conditions. In simple terms, users must understand what models can and cannot do, where they apply, and where they do not apply!

Generalized Hydrologic Simulation Models

Hydrologic simulation models are tools for addressing questions related to the analysis of past, present, and future streamflow regimes, changes of hydrologic processes, and issues of hydrologic design. These models are useful in extending hydrologic and watershed data and other information for predicting what might happen when something has changed on a watershed landscape. Hydrologic simulation models also represent a framework for inquiry that helps researchers to focus on meaningful problems and prioritize their research. To properly apply any given simulation model, hydrologists need to understand how they were developed and for what purposes, as well as their spatial representation, temporal resolution, and conceptual basis.

Development

A hydrologic simulation model is a simplification of an actual hydrologic system. Models are constrained by our ability to mathematically represent the complexities of hydrologic processes, biological and physical functions, and the integrated hydrologic response of watersheds to various climatic inputs. We will stress mathematical models rather than physical models because they are more broadly used and generally applicable because of their representation of hydrologic processes on watersheds and their accounting of water storage components and conveyance systems over time (Larson, 1973; Dingman, 2002). These models are developed for

- formulating and testing hypotheses about hydrologic processes;
- estimating streamflow responses to watershed management practices and land-use activities for differing climates and precipitation events including those that fall outside the measured conditions.

The utility of these models is their “capacity” to generate reasonably accurate predictions within the context of the available input data. *Deterministic models* are the focus here because they are better suited to predict the hydrologic effects of a change in vegetation, land use, or climate than *stochastic models*. The criteria for classifying hydrologic simulation models include their spatial representations, temporal resolutions, and conceptual basis.

Spatial Representation

Hydrologic models are either *lumped* or *spatially distributed* in their formulation. Lumped models assume no spatial heterogeneity in the modeling domain and do not allow for spatially variable inputs. Spatially distributed models allow the user to designate varying precipitation, temperature, and other climatic variables and the spatial occurrence of vegetation types, soils, and slope. Hydrologic processes such as interception, infiltration, soil moisture content, and surface and subsurface flow are represented on a landscape in spatially distributed models as varying on a landscape.

Temporal Resolution

Hydrologic simulation models are either *single event* or *continuous simulators*. Single-event models simulate the response to a specific precipitation event while continuous simulation models predict streamflows during and between rainfall or snowmelt events. Such models keep track of antecedent moisture conditions of a watershed and are well suited to predict high and low streamflows and the corresponding sediment transport and other water-quality characteristics over periods of time. Hydrologic simulation models provide outputs of these characteristics, typically in an hourly, daily, or seasonal time steps or in steady-state or longer-term average conditions.

Conceptual Basis

Hydrologic simulation models range from *empirical* to *physically based* in their structure. Empirical models relate streamflow outputs to precipitation inputs without simulating the hydrologic processes that are involved (Larson, 1973) and are sometimes called "black box" models. The parameters in empirical models are often developed from a synthesis of the results of field experiments, other published or unpublished information and data sets, and previous modeling results. These parameters are adjusted when necessary by comparing simulation to observed outputs with adjustments made until the best fit is achieved.

Physically based models are the most "conceptually correct" models in which the mathematical relationships (algorithms) represent soil, vegetation, and topographic characteristics and precipitation–temperature inputs that are obtained from on-site measurements. Because there are limitations in mathematically representing all of the hydrologic processes of a watershed, physically based models are likely to contain empirical relationships that require a fitting or lumping of characteristics at some scale. Most models that are used operationally by agencies and consulting firms are more empirical than physically based.

Examples of Generalized Hydrologic Simulation Models

The development of hydrologic models has been, and continues to be, an iterative process for the most part. The result of this process has been an assemblage of hydrologic simulation models in which there is no single tool that is suitable for all possible tasks. A description of hydrologic models that are used operationally in the USA is beyond the scope of this book; however, the reader is directed to Singh and Frevert (2009) and Singh (1995) for a more thorough coverage. Although each model differs in their methods of simulating streamflow, most generalized continuous simulation watershed models have the general form illustrated in Figure 16.1.

Several operational models are derivations of the Stanford Watershed Model (Crawford and Linsley, 1966) and include the Hydrological Simulation Program FORTRAN (HSPF) described by Donigan and Imhoff (2006) and models used for streamflow forecasting by the US National Weather Service (see the website for NWS River Forecast Centers: http://water.weather.gov/ahps/rfc/rfc.php). Additional models are the Precipitation Runoff Modeling System (PRMS) that represent a watershed as a series of hydrologic-response units (Leavesley et al., 1983, 1996) and the TOPMODEL that is based on knowledge of the spatial distribution of contributing areas and slope configurations of a watershed to simulate streamflow responses to precipitation events (Beven and Kirkby, 1979). More

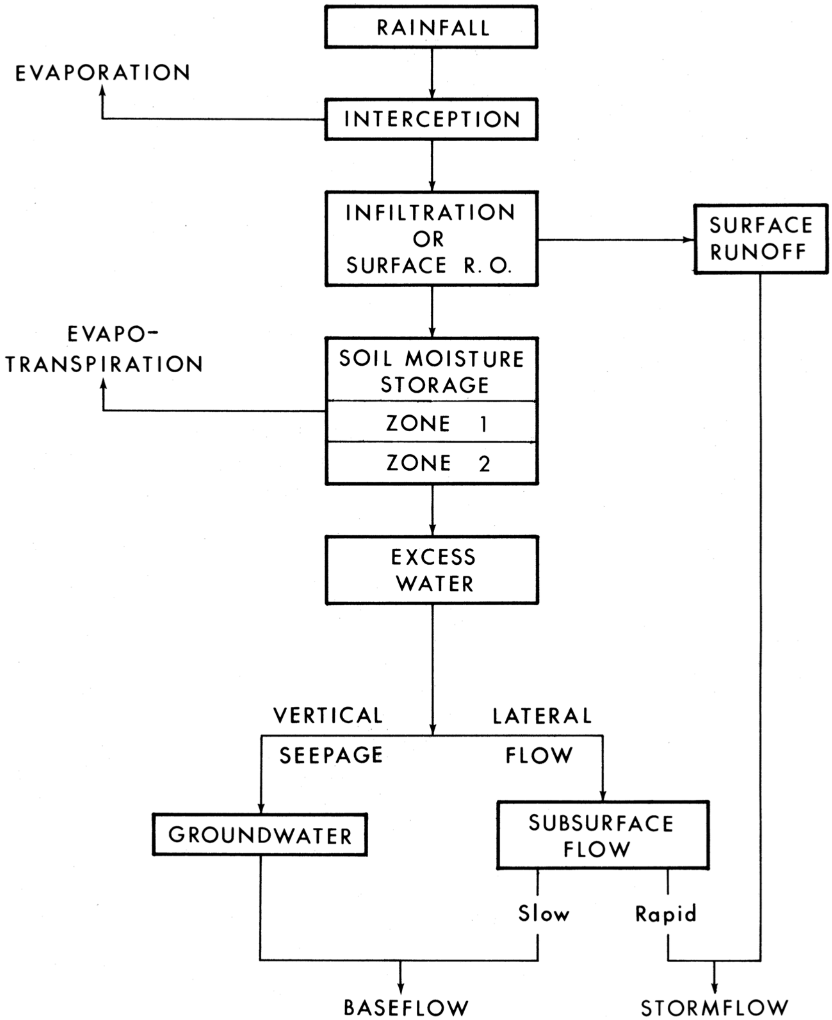


FIGURE 16.1. Hydrologic processes and runoff relationships commonly found in generalized continuous simulation models

recent models are process-oriented, spatially explicit, and physically based structures such as MIKE SHE (Refsgaard and Storm, 1995), DHSVM (Wigmosta et al., 2006), and Gridded Surface/Subsurface Hydrologic Analysis (GSSHA) (Downer et al., 2006). These models utilize geographic-referenced technologies and geographic information systems (see the following sections) to simulate the hydrologic response to changes.

Applications of Generalized Hydrologic Simulation Models

Hydrologic simulation models must be able to facilitate the scaling-up of hydrologic predictions from small to large watersheds and to larger river basins while examining long-term scenarios of hydrologic response to changing land-use activities. This requirement calls for models that are more physically based than empirical with the capacity to address changes within a framework of cumulative watershed effects.

Field studies that explain the spatial and temporal relationships of hydrologic processes are the foundation for the more physically based hydrologic simulation models. However, these models have often been restricted to research applications because of their extensive data requirements to operate. This poses a problem for watershed managers who are frequency constrained by availability of the required input data and the inherent complexity of the model. The challenge, therefore, has been to develop hydrologic models that are sufficiently physically based to produce reliable simulations but can be operated with data that are commonly available. Such models must be capable of

- simulating hydrologic responses to change on small- and large-scale watersheds in a timely manner with input data that are readily available;
- predicting streamflow and water-quality responses to changes on watersheds representing a mosaic of land-use conditions;
- representing key hydrologic processes mathematically with acceptable accuracy to provide the right answer for the right reason; and
- simulating responses to changes for a range of climatic and land-use conditions for which there is uncertainty about future management options.

A feedback loop connecting research, modeling, and management frames the development of hydrologic simulation models. An iterative process is often followed to integrate the development of hydrologic models for applications on watersheds. In this process, inaccuracies in the modeling results should be analyzed to determine deficiencies in the algorithms that might not adequately represent hydrologic processes. The models can be modified and tested where hydrologic-process data are available at the watershed scale.

Applications of Generalized Hydrologic Simulation Models on Forested Watersheds

Hydrologic simulation models that are used operationally must often be capable of simulating streamflow from forested watersheds and in some cases they have been used to simulate responses to changes in the forest cover on a watershed. Examples of these models are HSPF (USEPA, 1984) and the Soil and Water Assessment Tool (SWAT) (Neitsch et al., 2002). While these models have been applied to forested watersheds, they were not developed to simulate forest effects on streamflows. Therefore, their applications are limited to situations where they can be calibrated with data sets corresponding to the forest conditions for which they are applied.

The reliability of models based on "linear relationships" between precipitation and surface runoff or other relationships derived for systems that produce overland flows of water are not always well suited for use on forested watersheds (Eisenbies et al., 2007). For example, the curve-number method (see Chapter 8) does not adequately represent streamflow processes on forested watersheds (Hawkins, 1993; Hawkins and Khojeini, 2000). Similarly, the Revised Universal Soil Loss Equation (Renard et al., 1997) that estimates soil erosion from rain splash, sheet wash, and rill erosion are based largely on overland flow processes and, therefore, are not applicable to most forested watersheds. It is necessary that users of these models be aware of these limitations in simulating stormflows and erosion-sediment processes on forested watersheds. It is often better to use hydrologic models that have been developed to simulate streamflow response from forested watersheds.

Forest Hydrology Simulation Models

Research efforts have led to model development that has departed from hydrologic simulation models of the past in which surface runoff was the dominant processes in generating streamflow responses. It has been found, for example, that subsurface flow can be the "dominant mechanism" in generating streamflow from many forested watersheds in humid regions with deep soils (see Chapter 5). Furthermore, hillslope processes in such forested watersheds generate stormflow following the variable source area concept, which has been depicted by Hewlett and Troendle (1975) and described by the model of Troendle (1985). As a consequence, a shift has occurred in the structure of forest hydrology models from lumped-parameter models to those representing hydrologic processes that vary with the landscape position within a forested watershed.

Desirable Characteristics

Desirable characteristics of forest-hydrology simulation models include capabilities of

- simulating hydrologic responses to forest management practices or changes in land use on forested watersheds;
- accounting for the spatial distributions of watershed management practices or land-use changes and depicting the hydrologic responses through changes in hydrologic processes;
- simulating the hydrologic processes of interception, infiltration, transpiration, surface runoff, subsurface flow (interflow), soil moisture and groundwater–surface water exchanges of forested watersheds; and
- providing continuous simulations for a wide range of vegetative, physiographic features and land-use conditions in which forest cover is a component.

Among the criteria for evaluating and then selecting among the available models for applications on forested watersheds are

- Accuracy of prediction – Models with minimal bias and error variance are superior.
- Simplicity – Refers to the number of parameters that must be estimated and the ease with which the model can be explained to the user.
- Consistency of parameter estimates – This requirement is important in models that use input parameters estimated by optimization techniques. Models are unreliable when optimal values of the parameters are "sensitive" to the period of record in the simulation exercise or the values vary widely among similar watershed conditions.
- Sensitivity of results to changes in parameter values – It is desirable that models not be sensitive to input variables that are difficult to measure and costly to obtain.

Examples

A review of forest-hydrology simulation models suggests that a few have gained widespread application. Examples are presented to illustrate their application and constraints in their use.

The *Water Resources Evaluation of Nonpoint Silvicultural Sources* (WRENSS) handbook (USEPA, 1980) describes models recommended in 1980 for simulating the effects of watershed management practices and land-use changes on water yields, sedimentation, and water quality. Models were recommended for rain- and snow-dominated regimes to provide managers with tools that are process-based and rely on data sets that are readily available.

Models developed largely for rainfall-runoff regimes include TOPMODEL, BROOK (Federer and Lash, 1978; Federer, 1979), PRMS/MMS (Leavesley et al., 1983; 1996), and DHSVM (Wigmosta et al., 1994; 2006). PROSPER (Goldstein et al., 1974) is a soil–plant–atmosphere model recommended for use in rain-dominated catchments. However, application of PROSPER requires detailed knowledge of soil–plant–atmosphere characteristics that might not be available in many situations. Other rainfall-runoff models have been applied on forested watersheds to evaluate the effects of timber harvesting and silvicultural cuttings (Putz et al., 2003; Waichler et al., 2005) and road networks (LaMarche and Lettenmaier, 2001; Tague and Band, 2001; Cuo et al., 2006) on streamflow regimes.

A model capable of simulating streamflow in snow-dominated regions is the Subalpine Water Balance Model (WATBAL) originally formulated by Leaf and Brink (1975). Later versions of WATBAL incorporated remote-sensing data as input variables. Initially verified for operational applications in the Rocky Mountains of Colorado, versions of the model are currently applied throughout the Mountain West. WATBAL is a lumped-parameter continuous simulator formulated to simulate the effects of watershed management practices and land-use changes on a range of hydrologic components including streamflow response to snowmelt-runoff, rainfall, and a combination of these two streamflow generators.

Technologically Advanced Tools

By the time this book goes to print, some of these applications may be replaced by a new emergent technology currently unknown to the authors. Nevertheless, we present information about tools that help us understand the landscape better, which in-turn, offers information to watershed managers that may improve predictive power for wise land-use management. Tools discussed include remote sensing, radar, GIS, GPS, LiDAR, and others.

Remote-Sensing Platforms

Many remote-sensing platforms are operational, others have completed their usefulness, and still others are currently in design or in early production phases (Guertin et al., 2000). There are many types of imaging applied in earth-science observations, modeling, and the management of land, water, and other natural resources including visual, ground- and aerial-based photography, satellite observations, radar, and sonar. While significant advances have been made in nearly all phases of remote-sensing technologies, the emphasis has often been placed on the electromagnetic (EM) spectrum. Two types of EM sensing most commonly used in studying the features on watershed landscapes are optical remote sensing and radar.

Optimal Remote Sensing

Optical remote sensing focuses on the short wavelengths of the EM spectrum from ultraviolet to the infrared spectra. Optimal remote sensing has advantages over "traditional" land-surface studies that have been limited by point-based estimates of the parameters of interest and constrained by sample size. It provides a unique and useful perspective of the earth's surface and it is also suitable for large-scale investigations of surface patterns. Many mathematical relationships are available for case-specific applications in classifying vegetation, geomorphology, and soil analysis (Guertin et al., 2000; NRC, 2008). One such algorithm is the *normalized difference vegetation index* or NDVI that is a spectral ratio between the infrared and red spectra to predict biomass. However, there are also constraints to the use of optimal remote sensing including operational limitations to daylight hours and the masking of surface signals from the sensors because of clouds and smoke.

Radar

Radar (radio detection and ranging) uses the microwave (long-wave) portion of the EM spectrum that is orders of magnitude longer than those sensed in the optical range. Most radar applications involve the emission of a microwave signal from an aircraft or satellite toward the object of interest with the aircraft or satellite recording the signal upon its return. Synthetic aperture radar (SAR) facilitates detailed mapping of surface characteristics through processing radar signals such that the azimuth resolution is improved in proportion to the system's aperture size (Dobson, 2000). Interferometric SAR, where a target is sensed multiple times from different orientations, can be used to prepare highly detailed maps that detect changes in the surface of a watershed, flood events, and tree-harvesting operations, and the effectiveness of large-scale watershed management programs.

Geographic Information Systems

Geographic information systems (GIS) are capable of capturing, storing, analyzing, and retrieving geographically references data on demand in a format that meets the informational needs of a hydrologist or watershed manager (Star and Estes, 1990; Guertin et al., 2000). GIS data represent objects including waterways, land use, trees, roads, and elevations with digital data stored (traditionally) in either a *raster* or a *vector* form. The data layer is represented by an array of rectangular or square cells each of which has an assigned value in a raster (cell-based) system while the line work in a vector (line-based) system is represented by a set of connected points with a line segment between two points considered a vector. The choice of data layers is dependent on the needs of the user (Fig. 16.2).

Applications

GISs often together with computer simulation models and remote sensing techniques have applications in hydrology and watershed management including keeping current inventory information such as the quantities of a resource available, where it is located, and whether it is growing, shrinking, or stable in character (Benda et al., 2007). Incorporating information from a GIS into hydrologic simulation models provides an element that other hydrologic models often lack, that is, an ability to analyze combinations of slope, aspect, and hydrologic-response units into the simulations. More detailed information can also be included in the models such as vegetation types, terrain roughness, and soil characteristics that can influence infiltration and transpiration rates and, therefore, surface runoff (Heywood et al., 2006). These added data layers can result in more accurate models.

GISs make possible collaborations among watershed researchers, managers, policy-setters, and stakeholders in the decision-making process. The time of managers and researchers preparing and then presenting management alternatives to the public without

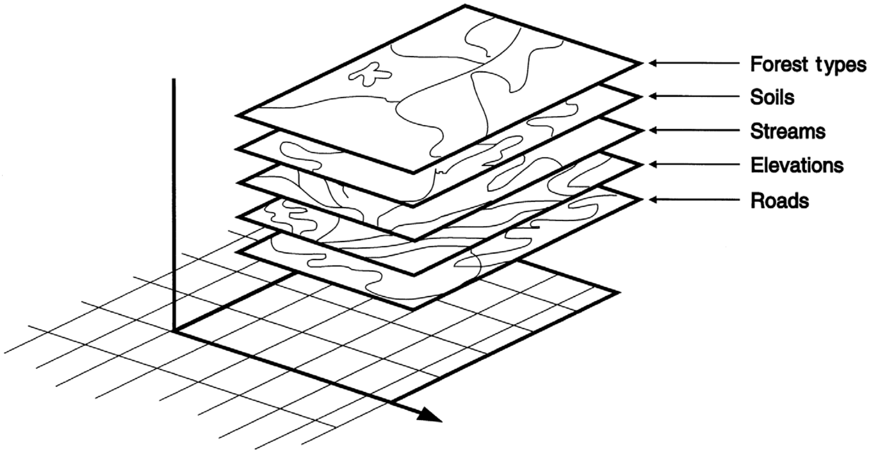


FIGURE 16.2. A GIS conceptualized as a set of geographically registered data layers

incorporating current science, social perspectives, and economic interests has passed. Professionals are participants and facilitators in the decision-making process with possibilities of GIS playing a significant role in this regard.

Possible Errors

The accuracy of a GIS depends on the source data and how it is encoded into the system. The spatial data sets for a GIS are obtained largely from maps, aerial photographs, satellite imagery, and traditional and global positioning surveys. Each source of data involves a number of transformational steps and transformations from the original measurements to the digital coordinates each of which can represent a potential error in the system (Bolstad and Smith, 1992). Inaccuracies of geographical data are common and are propagated throughout the applications of a GIS in ways that are difficult to predict. Therefore, the user must recognize that the outputs of a GIS are not necessarily error free.

Global Positioning System

The global positioning system (GPS) is a space-based global-navigational satellite system that provides location and time information on weather almost everywhere on, or near, the earth where there is an unobstructed line of sight to four or more GPS satellites. It is maintained by the USA government but is freely accessible to anyone with a GPS receiver. While originally a military project, GPS also has civilian applications. For example, GPS technology is advancing surveying by providing "absolute locations" to determine boundaries and make maps. The location coordinates obtained by GPS can be linked to digital objects such as photographs and other documents for creating map overlays for inclusion in a GIS when appropriate.

Applications

A GPS satellite can fix the locations of flumes and weirs, weather stations and remote precipitation gauges and other monitoring devices, watershed-research study plots, and measured cross sections of stream channels. GPS-referenced multiple-resource data can also be incorporated into database management systems to facilitate interpretations of watershed characteristics in planning and managing a land and water resources (Habraken, 2000). These characteristics can be watershed descriptors such as its designation of the watershed, size, orientation, streamflow network and physiographic characteristics, streamflow regimes, soil erosion and sedimentation, tree overstories, understories of herbaceous plants, loadings of flammable fuels, and wildlife habitats. GPS satellites are limited in their ability to fully characterize the spatially distributed characteristics of a watershed, however.

Errors

Errors in GPS applications occur. These errors are affected mostly by geometric dilution of precision and depend largely on atmospheric effects, signal arrival-time errors, numerical errors, ephemeris errors, and multipath errors.

Decision-Support Systems

Watershed managers are often faced with the task of making appropriate, responsible, and acceptable decisions on the use of land, water, and other natural resources. The complexity of the questions asked by people, the extensive information available, and the often decreasing availability of land, water, and other natural resources are why better decision-making procedures have become necessary. Fortunately, a diversity of decision-support systems has become available to help managers make better decisions about watershed management (Lane and Nichols, 2000). Linear programming and multiple-criteria decision-making techniques are two of these methods.

Linear Programming

Linear programming can be the basis for relatively simple decision-support systems. For example, a watershed manager might wish to reduce the costs of a management practice represented by a linear-objective function and a set of linear constraints. However, the one and only one objective function and all of the constraints must be linear for the range of values permissible. Unfortunately, this basic requirement of linearly is inappropriate or unsuitable for many watershed management problems.

Multiple-Criteria Decision-Making Techniques

Decision-making problems more commonly involve several objectives that are not be linear. Reducing the costs of management while simultaneously optimizing the total benefits obtained from soil and water conservation practices, wood production, and wildlife-habitat protection might be necessary. Such a set of objective functions are likely to be subject to several linear and nonlinear constraints. More realistic problems such as this can be analyzed by multiple-criteria decision-making (MCDM) techniques. MCDM techniques determine the preferred solutions to management-related problems in which the discrete alternatives are evaluated against specified acceptance criteria.

It is beyond the scope of this chapter to outline the details of the many MCDM techniques available to a user. The reader is referred to Yakowitz et al. (1993), Anderson et al. (1994), El-Swaify and Yakowitz (1998), and others for further information on these techniques.

Internet Applications

The internet is a network of networks linking computers to computers. The internet itself does not contain information but it provides users with access to information that is available on a worldwide basis. It would be more correct, therefore, to say that the information was *found* by using the internet. Internet communications used most frequently by people is email, which is an easy-to-use way of delivering content and receiving feedback. A collection of email addresses allows a user to deliver messages of interest to many people with a similar interest on a Listservs® named for the software used for this purpose.

The world wide web (the web) is the largest and fastest growing activity on the internet and has been referred to extensively in this book. While incorporating all of the internet services available, the web is also a system of servers that support formatted documents with links to other documents with text, graphics, audio, and video files. Users can find publications, data sets, images, bibliographies, and software about hydrology and watershed management on many web listings. Some of these listings have been referenced throughout this book. Subject guides and search engines are available to help people to effectively collect and distribute this information. Another way of obtaining this information is attaching website addresses to communications with colleagues.

The Teaching Library at the University of California, Berkeley, offers free drop-in classes on the internet, the web, and finding information by using the internet. Access http://www.lib.berkeley.edu/TeachingLib/Guides/Internet/ for further information.

Remote-Sensing Technologies

Airborne and satellite remote-sensing techniques and light detection and ranging (LiDAR) are two remote-sensing technologies that continue to emerge in development and application.

Airborne and Satellite Remote-Sensing Techniques

Airborne and satellite remote-sensing technologies have improved to the time and costs of monitoring the storage of water in the atmosphere, changes in vegetation and soil characteristics, and importantly ET processes on large watersheds, river basins, and geographic regions (NRC, 2008). Recently launched satellites such as Terra and Aqua that carry the Moderate Resolution Imaging Spectroradiometer (MODIS) sensor are viewing the surface of the earth every 1 or 2 days to acquire source-data in 36 spectral bands. Daily estimates of ET losses at 1 km (0.6 mi) resolution are available with MODIS imagery. MODIS with scaling techniques to reconcile differences in resolution is also used to provide water-balance information at fine spatial and temporal scales (Singh et al., 2004). Other satellites operating in the optical and microwave parts of the spectrum are useful in mapping areas of inundation and saturation on boreal landscapes (Sass and Creed, 2007) and wetland ecosystems (Toyra et al., 2001).

Light Detection and Ranging

Light detection and ranging (LiDAR) is an optical remote-sensing technology that measures the properties of scattered light to facilitate the analyses of topographic features, land-use characteristics of watersheds, and the climatic and precipitation regimes effecting a watershed (Lefsky et al., 2002, and others). The primary difference between LiDAR and radar is that LiDAR uses shorter wavelengths (ultraviolet, visible, or near infrared) in the EM spectrum than radar. LiDAR is highly sensitive to cloud particles and, therefore, has applications in atmospheric investigations, meteorology, geomorphology, and seismology.

High-resolution digital maps that can be generated by stationary and airborne LiDAR have also led to advances in hydrology and watershed management. For example, scientists from the National Oceanic and Atmospheric Administration (NOAA) and National Aeronautics and Space Administration (NASA) have been able to study changes in streambanks and shorelines. The topographic information obtained by LiDAR has been used for sitting stream-gauging networks on remote watersheds with dense tree overstories (Poff et al., 2008). LiDAR can also provide details of forest structures on watershed landscapes that are comparable to the information obtained from traditional field inventories but on a much larger scale (Hummel et al., 2011). At this time, however, advances in applications of LiDAR appear to have outpaced the development of hydrologic models capable of representing hydrologic processes and watershed responses associated with land-use conditions that range from forests and wetlands to agricultural croplands and urbanization.

Advanced Geographic Information Systems and Geovisualization

Advanced GISs when combined with spatial modeling methods have greatly enhanced the collection and analysis of large spatial and temporal data sets to predict water flows and sediment transport from small and large watersheds. For example, the NetMap system can be used to estimate hillslope failures, soil-erosion potentials, and sediment supplies (Benda et al., 2007). GIS with models of terrain-analysis features can facilitate the planning for stream restoration. The results obtained from these applications can be useful in comparing the impacts of alternative watershed management practices while assessing the cumulative watershed effects of the practices.

These technologies continue to progress in allowing predictions of the effects of watershed management and land-use activities at finer resolutions and for larger areas than were possible only a decade ago.

Multisensor Networks

Development of multisensor networks connected through wireless technologies is underway. However, securing reliable power sources, the challenges of processing large amounts of data, and other engineering constraints must be overcome. Limited testing of these networks has been made to measure temperature and soil moisture at high spatial and temporal resolutions on plots or small watersheds (NRC, 2008). Multisensor networks have not been implemented on larger spatial scales at this time, but hold promise to improve our understanding of hydrologic processes at finer scales.

Using the Stable Isotopes of Hydrogen and Oxygen

Isotopic tracers are useful in studying hydrologic processes. Where surface water and groundwater interact, such as the case in riparian areas, lakes and wetlands, it is often of interest to be able to determine the flow pathways and contributions of groundwater and surface water to a body of water. If one wanted to restore or create a wetland, the contributions of groundwater and surface water would be of importance in determining the type of wetland that could be restored or constructed. Stable isotopes of oxygen and hydrogen can be used to determine the contribution of water from rainfall runoff (new water) in contrast to the contributions of water that has had significant residence time in the soil and/or groundwater aquifers (old water). This distinction becomes important when designing treatment BMPs or selecting an appropriate water resource protection plan.

An *isotope* is an element with a specific number of neutrons and protons in its nucleus. The isotopes ^{18}O, ^{2}H, and ^{3}H are components of natural water molecules that fall to earth as precipitation and, therefore, are useful tracers for watershed analyses. Hydrogen has two stable isotopes, a lighter isotope protium or ^{1}H and a heavier isotope, deuterium (D), or ^{2}H. There is also a third isotope, tritium, or ^{3}H, which is radioactive and has a half-life of 12.3 years. About 99.985% of the hydrogen atoms in the hydrosphere comprise ^{1}H, while ^{2}H only accounts for 0.015%. Oxygen has three stable isotopes: ^{16}O comprises 99.859%, ^{17}O 0.038%, and ^{18}O 0.2% of the oxygen atoms in the hydrosphere. The heavier isotopes represent a small fraction of the total, but this difference can be measured and the ratio between the lighter and heavier isotope is determined. An agreed-upon standard that has been set by the International Atomic Energy Agency, known as the Vienna Standard Mean Ocean Water (V-SMOW), is used to report the differences between the ratio of the heavy isotope to the lighter isotope. Values of δ are reported in units of parts per thousand (reported as per mill or ‰) relative to a standard with a known composition (Kendall and Caldwell, 1998) as

$$\delta(‰) = (R_x/R_s - 1) * 1000 \qquad \textbf{(16.1)}$$

where R is the ratio of the heavy isotope to light isotope, R_x is the ratio of the sample, and R_s is the ratio of the standard.

Processes of evaporation and condensation can produce variations in the ratio between lighter and heavier isotopes. For example, when water evaporates from surface water, molecules containing heavier isotopes evaporate at a slightly slower rate than the molecules containing lighter isotopes. During a thunderstorm, molecules containing heavier isotopes will tend to fall more readily in the precipitation than molecules containing lighter isotopes. These processes redistribute the relative abundance of δD and $\delta^{18}O$ as water moves through the hydrologic cycle. The term used to describe this redistribution is known as *fractionation*.

Isotopes have been used in watershed studies to estimate evaporation, groundwater recharge areas, groundwater age, relative hydraulic residence time, hydrograph separation in runoff studies, and characterization of groundwater–surface water interactions. The isotopic composition of water in deep and shallow aquifers suggests different recharge source areas. Elevation difference between plains and mountains produce isotopic variation because mean annual temperature changes with elevation. Magner et al. (2001) observed isotopic compositions that were similar to snow for the region in a confined outwash aquifer below the Red River near Moorhead, Minnesota. Additional carbon dating confirmed the age of the aquifer water to be of glacial origin. Without the isotope information, recognition of ice age water and future management ramifications of withdrawal would have been less clear. The application of stable isotopes were successful largely because of unique hydrologic, geologic, and landscape features of the study area. Stable isotopes are commonly used in hydrology for estimating source waters and hydrologic pathways and hydraulic residence time.

Burns and McDonnell (1998) estimated hydraulic residence time by comparing the amplitude of a best-fit isotopic sine curve for precipitation to the amplitude of a similar curve for water that is of interest. Seasonal changes in the $\delta^{18}O$ composition of precipitation at temperate latitudes tend to follow a sinusoidal pattern. The pattern occurs over a period of 1 year, reflecting the seasonal changes in tropospheric temperature. For a given location, measured changes in $\delta^{18}O$ composition for a water body of interest (e.g., lake, pond, stream, soil water, or groundwater) are made during different seasons. Mean hydraulic residence can be calculated, if the seasonal waters are considered in a steady-state, well-mixed reservoir with an exponential distribution of residence times as follows:

$$T = \grave{u}^{-1}\,[(A/B)^2 - 1]^{1/2} \tag{16.2}$$

where T is the estimated hydraulic residence time, in days, $\grave{u}$ is the angular frequency of variation (2π/365 days) or (0.0172), A is the input amplitude, and B is the output amplitude. The input amplitude can vary year to year depending on predominant temperature of the precipitation. For example, in Minnesota winters are typically cold with precipitation coming from Arctic sources yielding $\delta^{18}O$ values in the −20s. Convective storms in the summer months, arising from Gulf-stream air (Gulf of Mexico), can be relatively heavy ($\delta^{18}O$ values: −2 to −5 range). These extremes in temperature suggest A could range from 15‰ to 20‰. Climate change could be decreasing the range because more winter precipitation is coming from the Gulf air stream, coupled with Arctic air producing cooler summers. Even with climate change, the output amplitude B should concordantly respond to seasonal temperature shifts. For example in Table 16.1 a range of output amplitudes for Ten Mile Lake in northern Minnesota is less than 1. Between some seasons B is near zero and other seasons B is approaching 0.1, which is reflective of year to year variation. Using numbers on either extreme will give erroneous hydraulic residence times.

Stable Isotopes and Minnesota Lakes

Engel (2012) sampled a series of lakes across Minnesota from 2008 through 2010 as part of a Sentinel Lake study lead by the Minnesota Department of Natural Resources (MDNR). Because lakes typically lose water via evaporation during the summer months, some insight

TABLE 16.1. Per mill change of $\delta^{18}O$ values collected from Ten Mile Lake, Minnesota, between May 21, 2008, and October 11, 2010. Data show high amplitude of 0.71 and a low amplitude of 0.01, with an average amplitude change of 0.3%

Sample date	$\delta^{18}O$ (‰)	% change
May 21, 2008	−4.15	
July 15, 2008	−4.39	0.24
October 7, 2008	−3.71	0.68
May 19, 2009	−4.42	0.71
July 14, 2009	−4.29	0.13
October 20, 2009	−4.25	0.04
May 19, 2010	−4.70	0.45
July 28, 2010	−4.71	0.01
October 11, 2010	−4.60	0.11

into regional lake storage and the unique hydrology of each sentinel lake was obtained by developing regression models of lake-isotope data. One of the ways to detect evaporation is by plotting $\delta^{18}O$ versus δD. because the deuterium will evaporate slightly faster than the heavy oxygen, evaporated waters will shift or plot to the left of the linear line of isotope data.

Dansgaard (1964) plotted $\delta^{18}O$ against δD for worldwide precipitation waters and found a linear line, defined as the global meteoric water line (GMWL). The data had and $r^2 = 0.997$ ($n = 74$ stations) where $\delta D = 8.17\delta^{18}O + 10.55$. This slope varies for different geographical areas depending on the differences in mean monthly air temperature but in general $\delta D = 8\delta^{18}O + 10$ is used to create a global MWL plot. For certain detailed studies, a local meteoric water line (LMWL) is required to detect evaporation; in Minnesota, Magner and Regan (1994) found $\delta D = 7.8\delta^{18}O + 9$, whereas in Québec, Canada, Sklash and Farvolden (1982) found $\delta D = 8.8\delta^{18}O + 22$. As previously discussed in Chapter 14, these lines may have changed over the last decade due to changing climate and precipitation patterns.

Figure 16.3 shows a relatively broad range of isotope data collected across Minnesota that plots to the left of the GMWL. The Canadian Shield data, collected near the border of Canada and Minnesota, consistently plot in the lower left containing lighter isotopes; whereas, the Prairie Corn-belt data collected in southern Minnesota, plot further to the upper right illustrating warmer input precipitation and more evaporative fractionation.

The stable isotopes of lake water $\delta^{18}O$ and δD plotted against each other create patterns associated with lake water residence time and movement over seasons. For example, data collected near the Canadian border in a lake carved by glacial scour of Canadian Shield bedrock with watershed water flowing through the lake tend to plot near the GMWL but

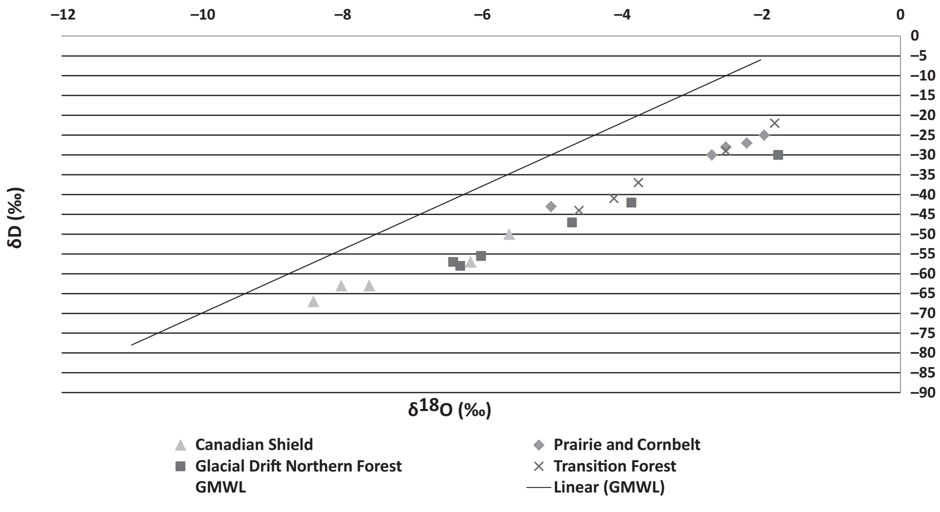


FIGURE 16.3. Isotope relationship of δD versus $\delta^{18}O$ (‰) for lakes across Minnesota; although the data overlap, a pattern of data is present based on northern and southern lakes; with southern lakes (Prairie and Cornbelt, Transitional Forests) plotting toward the upper right and northern lakes (Canadian Shield) plotting toward the lower left of the graph. The line is the L, global meteoric water line (GMWL).

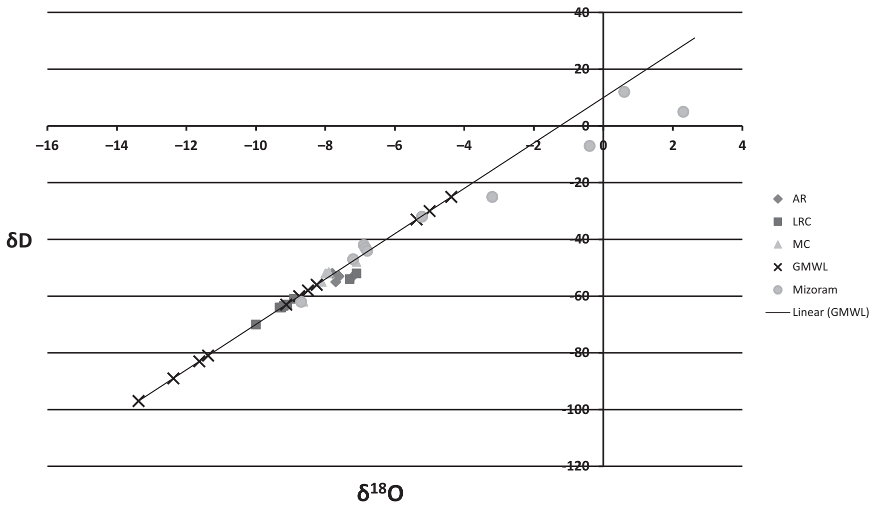
vary in location depending on season. In contrast, a lake that was formed by "ice contact" or a block of ice disconnected from its glacier surrounded by sand deposits will tend to plot to the left of the GMWL because very little watershed water is routed through the lake. Years of $\delta^{18}O$ data collected over multiple seasons (Table 16.1) will produce a sinusoidal pattern that can be used to estimate hydraulic residence time using Equation 16.2. Additional examples of the use of isotopes are illustrated in Boxes 16.1 and 16.2. This example and the others presented in Boxes 16.1 and 16.2 give just a sampling of how isotopes can be used in IWM. Clearly, understanding more about hydrologic pathways and processes can aid a watershed manager in decision support for wise land use and water resource management. The following weblink will provide more information on the use of stable isotopes in hydrologic studies: http://wwwrcamnl.wr.usgs.gov/isoig/res/

Box 16.1

The Influence of Elevation on Stable Isotopes

Mizoram is one of the northeastern states of India located between Bangladesh on the west and Myanmar on the east. Mizoram contains a fair amount of elevation change across the country with the highest mountain peak at 2210 m (7100 ft) and the lowest lying valley at 21 m (69 ft). Mizoram is set in a hilly terrain with few narrow floodplains located along the river valley. All human settlements are established either at the top of hills or on the hillslopes were spring water resurges. The entire state benefits from the direct influence of the monsoon rains and receives adequate annual rainfall: 2500 mm/year on an average, which occurs for the most part between second half of May to the first half of October. During the monsoon period water is abundantly available; however, there is limited water during the dry season that normally ranges from January to early May.

Land-use management has altered the soil surface and removed forest floor litter, which reduced soil infiltration capacity. During the monsoon season, rainfall runs off the soil surface resulting in insufficient recharge to groundwater, which has caused many springs, streams, and rivers to dry immediately after the monsoon season. The base flow of streams and rivers is no longer sustained during the dry periods of the year which results in acute water scarcity during the off-monsoon periods that especially affects the rural poor. Consequently, rural people spend time fetching water to meet their basic needs. Furthermore, the Government of Mizoram incurs sizable sums of money every year for supplying drinking water via trucks, which is hauled several kilometers from the water source to villages. Mizoram needed to find a way to reverse the current trend. To address this problem the hydrologic pathways of water movement in Mizoram needed definition. Stable isotope data were collected by the Government of Mizoram to determine if unique end members of source water existed throughout the state.



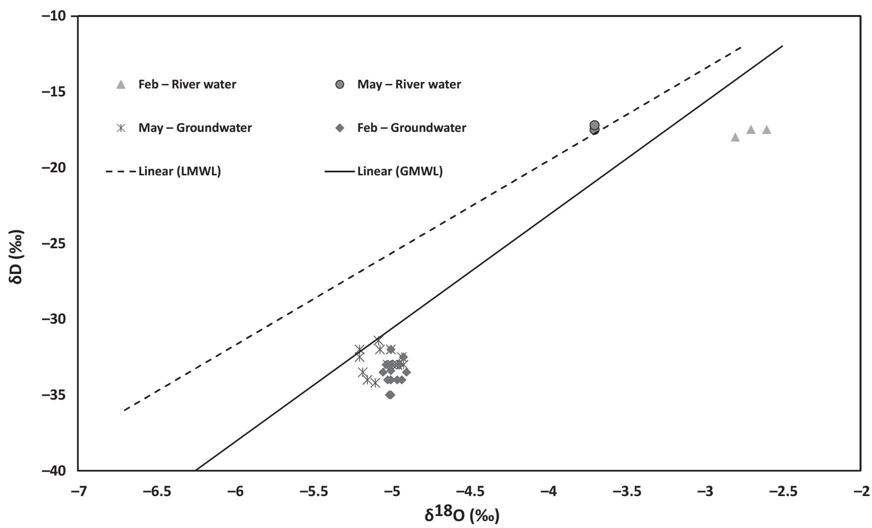
This Figure presents isotope data collected across Mizoram by elevation and by different types of water bodies: flowing water versus stagnant water (lake or pond). Minnesota data are also shown in the Figure to illustrate the contrasting strength of unique end members between the two different states; *end member* refers to a source of water with a unique signature that could mix with other source waters in a lake or river. The Figure shows relative clustering of Minnesota data on the GMWL (near $\delta^{18}O$ −9‰ and $\delta^{2}H$ −60‰) compared to the Mizoram data which extends to the upper right-hand corner beyond zero. Minnesota is relatively flat compared to Mizoram as described earlier; this contrasting elevation difference illustrates the power of elevation change, which in-turn, influences temperature. Isotopic values are strongly correlated with temperature as discussed earlier. Therefore, the stable isotopes of $\delta^{18}O$ and δD can be used as unique tracers of source waters in Mizoram to prioritize critical recharge zones that supply village springs.

Box 16.2

Groundwater and Surface Water Exchange

The Modder River near Bloemfontein, South Africa, is subject to high amounts of evaporation because of the climate and reservoir management. The Krugersdrif Dam creates an open-water area known as Soetdoring Nature Reserve. Evaporation from the open-water body produces a distinctly different isotopic signature from the regional infiltrated groundwater; the Nature Reserve water becomes enriched with heavy isotopes. Concerns about over-abstraction of local aquifer water for irrigation agricultural and reduced groundwater discharge into the river led to some isotopic analysis of source waters below the Krugerdrif Dam. If climate change is leading to less available water for aquatic life in streams and rivers in central South Africa, then laws governing the ecological limits of hydrologic alteration require changes in water use to accommodate minimum flows necessary to support aquatic life (Poff et al., 2010).

A plot of δ^{18}O against δD for groundwater and river-water samples below the Krugerdrif Dam is presented below. The May riverine samples plot on the LMWL, whereas the February samples plot to the left of both the LMWL and the GMWL; these data indicate a clear seasonal difference. The groundwater data, however, only vary slightly between seasons indicating a relatively long hydraulic residence time.



Deviation from the GMWL and the LMWL, GMWL: $\delta D = 8\delta^{18}O + 10$; LMWL is from Pretoria, South Africa: $\delta D = 6.5\delta^{18}O + 7.8$. (Note this LMWL is considerably different from the North America's LMWLs described above.) Based on samples collected in February and May of 2011, the groundwater δ^{18}O values ranged from −4.95‰ to −5.25‰ with a mean of −5.09‰. The groundwater δD ranges from −31.28‰ to −34.40‰ with a mean of −32.78‰. The mixed river water below the Krugerdrif Dam δ^{18}O values ranged from −2.70‰ to −3.64‰ with a mean of −3.12‰. The mixed river water below the Krugerdrif Dam δ^{2}H ranged from −15.15‰ to −17.70‰ with a mean of −16.63‰ (Gomo, 2012).

In October 2011 following a large (25-year) storm event, isotopes were collected at the Krugerdrif Dam outlet, 2 km below the dam outlet, and from adjacent groundwater monitoring wells along the 2 km river reach. In humid regions, the recession limb of the hydrograph is often dominated by groundwater discharge as previously discussed. An important management question related to aquatic health is "What was the likely percentage of groundwater mixed into the river below the dam outlet following the 25-year storm event?" The following simple mixing model provides one possible estimate based on the numbers shown earlier. (For this example we will use δD instead of δ^{18}O.)

$$\%\text{GW} - \text{Contribution} = (\delta_{\text{Rmix}} - \delta_{\text{Dam}}/\delta_{\text{GW}} - \delta_{\text{Dam}}) \times 100 \qquad \textbf{(16.3)}$$

where $Percent_{\text{GW-Contribution}}$ is the estimated groundwater contribution from subsurface water. δ_{Dam} is the δD of water measured at the reservoir outlet, δ_{GW} is the δD of groundwater measured from the adjacent monitoring wells, and δ_{Rmix} is the composite δD riverine mix below both the dam and known groundwater discharges. Although streamflow was not measured, an inferred estimate of the source water is given based on the isotopic composite measured in the stream below the defined 2 km mixing zone (R_{mix}). From the mean values of δD presented in the Figure, the groundwater δD was −32.78‰; the measured October 2011 value was −31.97‰, within the range presented in the Figure. However, the R_{mix} was not within the range described in the Figure (−15.15‰ to −17.70‰), but −6.17‰. Furthermore, water leaving the Krugerdrif Dam was relatively heavy (−3.92‰) compared to all of the isotope data collected in this project.

Using the October δD values discussed earlier, the estimated groundwater contribution would be $[6.17 - 3.92]/[31.97 - 3.92] \times 100 = 2.25/25.8 \times 100 = 8.7\%$ of the water collected at R_{mix} came from the regional groundwater discharge to the river. This percentage is likely low based on prestorm values shown in the Figure. We do not have prestorm dam outlet values; nevertheless by iteration of a range of possible dam δD, the groundwater contribution could be 10–30% higher than the calculated 8.7%. One possible explanation is the flushing of pre-event reservoir water into the floodplain and riparian corridor of the Modder River below the dam. As the river stage dropped, enriched heavy isotopic water flushed from the reservoir would displace the lighter, less enriched regional groundwater contained in the riparian corridor sediments. This process is illustrated in Figure 5.10d.

SUMMARY AND LEARNING POINTS

One of the major challenges facing watershed managers in the future is incorporating innovative applications of the available tools and emerging technologies into watershed management activities. More specifically, the challenge is how best to use these tools and the emerging technologies without impairing historical databases and compromising traditional but still acceptable methods of managing watershed landscapes (Stednick et al., 2004). The emerging technologies, described in this chapter, create new opportunities to test and better understand watershed processes. After completing this chapter, you should be able to

1. Discuss the way that the field and analytical tools described in this chapter are used by hydrologist and watershed managers.
2. Describe how the emerging technologies discussed in this chapter can help to improve a watershed manager's understanding of hydrologic processes and watershed values.
3. Describe the different types of hydrologic simulation models and explain
   - how empirical models differ from physically based models;
   - the attributes of continuous simulation models in contrast to single-event models; and
   - how lumped parameter models differ from spatially distributed models.
4. Explain the challenges of applying physically based hydrologic simulation models.
5. Explain the desirable capabilities of hydrologic simulation models capable of predicting land-use changes on watersheds.
6. List the desirable characteristics of forest hydrology models and indicate the constraints of applying generalized hydrologic simulation models to forested situations.
7. Describe how a watershed manager would use stable isotopes to solve difficult water use and land-use management issues.

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Appendix Units Commonly Used in Hydrologic Work, USA

Rainfall depths and rates

Rainfall depths are usually measured in inches or millimeters
Rainfall rates are usually reported as inches or millimeters per hour

Volumes of water

Cubic feet (ft^3)
Cubic meters (m^3)
Gallons
Acre-foot (acre-ft) – the volume of water required to cover one acre of land to a depth of one foot
Second-foot-day (sfd) – the volume of water accumulated by a flow of one cubic foot per second in a 24-hour period
Inches, millimeters, or centimeters (in, mm, or cm) – water volumes reported as depth measurements refer to the volume of water equivalent to the reported depth over the area of concern (usually a watershed area)

Discharge

Discharge is a volume of flow per unit time. Although discharge may be reported in any units of volume and time, the following are used most often:

Stream and river flows

Cubic feet per second (cfs)
Cubic meters per second (m^3/sec or cms)

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Flows used in agriculture or related to water storage

Acre-feet per unit time
Inches or centimeters per unit time
Acre-inches per hour (acre-in/hr)

Groundwater and municipal water supply flows

Gallons per minute, hour, or day (gpm, gph, gpd)
Millions of gallons per day (mgd)
Standard conversion factors from metric to English units

Quantity	Metric unit	English unit	To convert metric to English multiply by
Length	centimeters (cm)	inches (in)	0.394
	millimeters (mm)	inches (in)	0.0394
	meters (m)	feet (ft)	3.28
	meters (m)	yards (yd)	1.09
Area	square millimeters (mm^2)	square inches (in^2)	0.00155
	square meters (m^2)	square feet (ft^2)	10.76
	square meters (m^2)	square yards (yd^2)	1.196
	square meters (m^2)	acres	0.000247
	hectares (ha)	acres	2.47
	square kilometers (km^2)	square miles (mi^2)	0.386
Volume	cubic centimeters (cm^3)	cubic inches (in^3)	0.0610
	liters (L)	cubic feet (ft^3)	0.035315
	cubic meters (m^3)	cubic feet (ft^3)	35.3
	cubic meters (m^3)	cubic yards (yd^3)	1.31
	cubic meters (m^3)	acre-feet	0.000811
	liters (L)	pints	2.113376
	liters (L)	quarts	1.056688
	liters (L)	gallons	0.264174
Velocity	kilometers/hour (km/hr)	miles/hour (mi/hr)	0.621
	meters/second (m/sec)	feet/second (ft/sec)	3.28
Acceleration	meters/second^2 (m/sec^2)	feet/second^2 (ft/sec^2)	3.280839
Flow	cubic meters/second (m^3/sec)	cubic feet/second (ft^3/sec)	35.3
	liters/second (L/sec)	gallons/minute (gpm)	15.850322
Rates and yields	kilograms/hectare (kg/ha)	pounds/acre (lb/acre)	0.892183
	metric tons/hectare (t/ha)	short tons/acre	0.446091
	millimeters/hour (mm/hr)	inches/hour (in/hr)	0.03937
	centimeters/day (cm/day)	inches/day (in/day)	0.393701
Mass	grams (g)	ounces [avdp] (oz)	0.0353
	kilograms (kg)	pounds [avdp] (lb)	2.20
	metric tons (t)	short tons (ton)	1.10
Density	grams/cubic centimeter (g/cm^3)	pounds/cubic foot (lb/ft^3)	62.4
	kilograms/cubic meter (kg/m^3)	pounds/cubic foot (lb/ft^3)	0.0625
Force	newtons (N)	pounds force (lbf)	0.00986

(*Continued*)

Quantity	Metric unit	English unit	To convert metric to English multiply by
Pressure or stress	kilopascals (kPa)	atmosphere (standard) (atm)	0.00987
	kilopascals (kPa)	inches of mercury at 60°F	0.296134
	kilopascals (kPa)	millibars (mb)	10.0
	kilopascals (kPa)	feet of water at 30.2°F	0.33456
	kilopascals (kPa)	inches of water at 60°F	4.018655
	kilopascals (kPa)	pounds/square foot (lb/ft^2)	0.145038
	kilopascals (kPa)	pounds/square inch (lb/in^2)	20.885459
Temperature	degrees Celsius (°C)	degrees Fahrenheit (°F)	°F = (1.8°C) + 32
Energy	joules (J)	British thermal units (mean) (Btu)	0.00095
	joules (J)	calorie (cal)	0.239
	joules (J)	watt-hours	0.00028
Power	watts	foot-pounds/second (lbf/sec)	0.73756

Source: Adapted from American Society for Testing and Materials (ASTM), 1976, Standard for Metric Practice (Philadelphia: ASTM).

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