TWO

Wetland Soils, Hydrology, and Geomorphology

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The hydrology, soils, and watershed processes of a wetland all interact with vegetation and animals over time to create the dynamic physical template upon which a wetland's ecosystem is based (Fig. 2.1). With respect to many ecosystem processes, the physical factors defining a wetland environment at any particular time are often treated as independent variables, but in fact none of these variables are independent of the others. For example, the hydropattern of a wetland (the time series of water levels) is often considered a master variable that affects the soils, biogeochemistry, and biology of a wetland, but the hydropattern is in turn affected by the physical properties of the soil underlying the wetland. Any explanation of the physical factors defining the wetland template is therefore circular, and the order of presentation somewhat arbitrary.

Wetland soils are unique among soils. Soils of wetland environments possess physical, chemical, and morphological properties that readily distinguish them from upland soils. The accumulation or convergence of water in certain parts of the landscape alters the development, form, and chemical behavior of soils, creating a special class of soils known as *hydric soils*. The hydric soils in turn alter the movement of water and solutes through the wetland system. The soil is where many of the hydrologic and biogeochemical processes that influence wetland function and ecology occur. A complete understanding of wetland hydrology, wetland formation, wetland ecology, and wetland management requires a basic understanding of soils—including soil properties, soil processes, and soil variability—and of the hydrologic processes that control wetland systems. This chapter first introduces general principles and concepts of soil science, and then it describes the unique characteristics of wetland soils and explains how landscapes influence the local hydrologic cycle to lead to the development of wetland hydrology and wetland soils. Hydrologic concepts will be foreshadowed throughout the soils section.

Wetland hydrology, on the contrary, is not unique to wetlands. The commonly used phrase "wetland hydrology" should be considered shorthand for "hydrology, as it relates

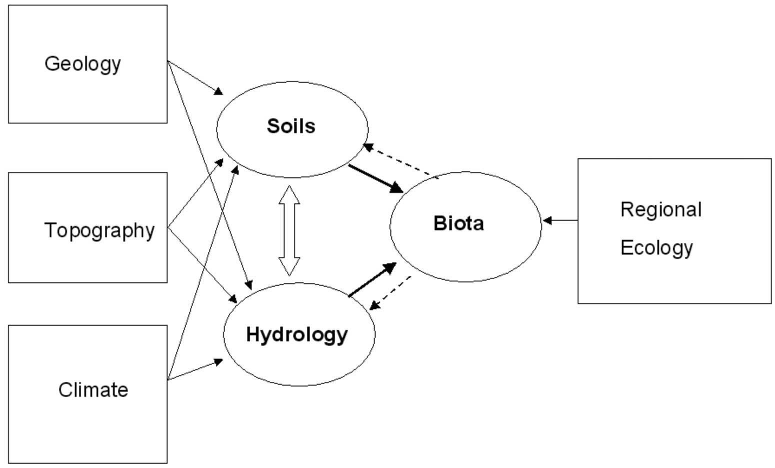

FIGURE 2.1. Abiotic and biotic influences upon and interactions within wetlands.

to wetlands" as there is no special sub-discipline of hydrology for wetlands. The principles and processes of hydrology can be applied to uplands, wetlands, streams, lakes, and groundwater. This chapter presents basic principles of hydrology that can applied to understand and explain annual, seasonal, and short-term water-level dynamics (the *hydropattern* or *hydroperiod*) of wetlands; to illuminate the physical and chemical water quality processes occurring in wetlands; and to explain the complex interactions among wetland hydrology, soils, and the resulting biotic community. Finally, the chapter presents broad ideas for understanding the geomorphic context for wetland formation and sedimentation.

Wetland Soils

Landscape Position

Wetlands occur where hydrologic conditions driven by climate, topography, geology, and soils cause surface saturation of sufficient duration to form hydric soils and competitively favor hydrophytic vegetation (Fig. 2.1). Landscape position is key to the processes affecting the formation and character of soils, as landscape position influences water movement over or through the soil and the local hydrologic budgets. Surface topography is a particularly important factor in controlling surface and subsurface water flow and accumulation. Although many landscapes are complex and irregular, there exist distinct and repeating patterns of hillslope elements that occur in most geomorphic settings. A typical hillslope profile (Fig. 2.2) can be segmented into summit, shoulder, backslope, footslope, and toeslope, landscape positions. The *summit* is the relatively flat area at the top of the slope. The *shoulder* is the steeply convex portion at the top of the slope. This surface shape favors the shedding of water and relatively drier soil conditions. The *backslope* is a linear portion of the slope that is not present in all hillslopes. At the bottom of the slope are the more concave *footslope* and *toeslope* positions, with the footslope being more steeply sloping than the toeslope. On such a typical hillslope, the quantity of water stored in the soils increases with proximity to the base of the hillslope in response to the accumulation of surface and subsurface flow from upslope positions. Concave contours promote convergent water flow, focusing surface and subsurface runoff to lower hillslope positions. Conversely, convex contours lead to divergent water flow. Across the landscape we can identify various landforms that represent different combinations of curvatures (Pennock et al., 1987), each of which affects the redistribution and storage of water. This, in turn, influences soil properties and wetland functions. Hillslope hydrologic processes and wetland water budgets are discussed in greater detail later in this chapter.

Soil Properties

Soils host the zone of biogeochemical activity where plants, animals, and microorganisms interact with the hydrologic cycle and other elemental cycles. A typical soil contains both mineral and organic materials as well as the adjacent water-filled and air-filled pore space. The physical and chemical properties of a soil may influence the processes that lead to wetland formation and function. Furthermore, wetland formation and function may influence some of the physical and chemical properties of soils. Biogeochemical processes in seasonally saturated soils can lead to the accumulation of organic matter and transformations of iron-based minerals, which may influence nutrient cycling, soil acidity, and soil color. Important soil physical properties include soil texture, soil structure, bulk density, porosity, and pore size distribution. These directly affect hydraulic conductivity, water storage, and water availability.

Color is the most apparent soil morphological property and it often indicates a great deal about the composition and the hydrologic conditions of a soil. Soil scientists quantify soil color using the Munsell color system, which uses three quantities—hue, value, and chroma—to define

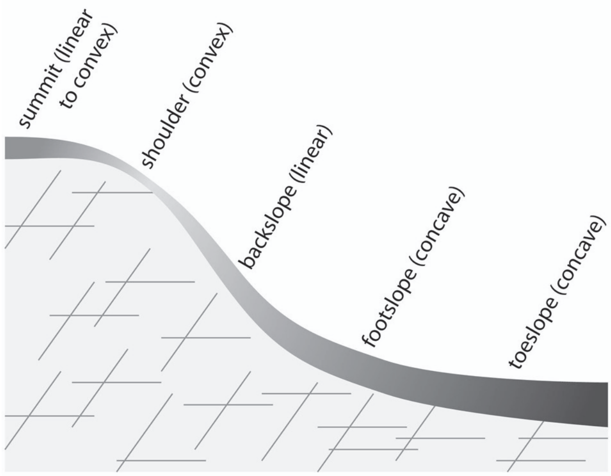

FIGURE 2.2. Typical hillslope cross section illustrating landscape positions of summit, shoulder, backslope, footslope, and toeslope.

a color. *Hue* refers to the spectral color: red (R), yellow (Y), green (G), blue (B), purple (P), or neutral (N), which has no hue. These five principal hues (not including neutral), plus the intermediate hues such as yellow-red (YR) or blue-green (BG), are used in the Munsell notation, with numbers placed before the hue letter(s) to designate four subdivisions within each of the 10 major hues. A progression of Munsell hues from reds to yellows is:

5R–7.5R–10R–2.5YR–5YR–7.5YR–10YR–2.5Y–5Y.

Value is a number between zero and ten that indicates the lightness or darkness of color relative to a neutral gray scale. A value of zero is pure black and a value of ten is pure white. *Chroma* is a number that designates the purity or saturation of the color. A high chroma indicates a pure color, meaning that there is one clearly dominant hue. A low chroma indicates that the color is a mixture of more than one hue. This is often illustrated when a small child uses a set of watercolor paints and, invariably, does not rinse the brush when changing colors. The resulting dull, drab, and muted colors have low chroma. For soils, chroma can range from zero to about eight. The format for writing a Munsell color is "hue value/chroma"—such as 10YR5/3.

Iron oxides and organic matter are the two primary coloring agents within most soils. Iron oxides give the soil a red, orange, or yellow color. Consequently, most soils are yellow-red in hue. Organic matter makes the soil brown or black, resulting in a Munsell color with a low value and a low chroma. The majority of the soil, though, is made up of aluminosilicate minerals, which are white to gray in color. In the absence of iron oxides or organic matter, soil color is dull and grayish with a high value and a low chroma. Such gray colors may be observed because iron oxides were never present in the soil. More commonly, gray colors arise because iron oxides have been reduced, became soluble, and translocated within the soil, usually because of saturated and anaerobic conditions. Uniform low-chroma colors are typical of prolonged saturated and anaerobic conditions in the soil. A mottled color pattern is often seen in soils that are wet for part of the year. The alternating patterns of red (high chroma) and gray (low chroma) colors indicate that some of the iron has been reduced or depleted (exposing the gray colors) in some areas, and has been concentrated in the red patches.

The mineral fraction of a soil contains particles of various sizes. Clay particles are, by definition, those that are smaller than 0.002 mm in diameter. Silt particles are greater than 0.002 mm but less than 0.05 mm. The largest soil particles are sand particles, which are greater than 0.05 mm but less than 2.0 mm. Any particles greater than 2.0 mm are collectively termed *rock fragments*. The most important aspect of soil particle size is the influence it has on surface area and pore size distribution. Clay particles have a high specific surface area, or surface area per gram of soil (up to 8,000,000 $cm^2\ g^{-1}$), whereas the larger sand particles have a low specific surface area (<1,000 $cm^2\ g^{-1}$). Most of the soil biogeochemical reactions occur at these particle surfaces, so soils with greater clay contents tend to be much more reactive.

The relative proportions of sand, silt, and clay particles determine the texture of a soil. For convenience, soil scientists have defined 12 different soil textural classes that cover various ranges in sand, silt, and clay content (Fig. 2.3). Soil texture influences almost every other property of a soil. Additionally, soil texture is a relatively stable soil property that does not readily change over time or in response to soil management. Soils within each textural class possess many similar characteristics and can be treated and managed in the same way.

In most soils the individual sand, silt, and clay particles are aggregated together to form secondary soil particles, or *peds*. The peds give the dry soil stability, and the spaces between the peds form the larger pores that promote faster water movement, greater gas exchange, and easier root penetration. The combination of texture and structure control the bulk density, porosity, and pore size distribution of a soil. The *bulk density* is the mass of dry soil per total volume of soil in units of g/cm^3 (see Equation 2.1), and it is inversely related to the total porosity (see Equation 2.2), which is the volume of pores per total volume of soil. In general, a sandy soil has a higher bulk density and a lower porosity than a clayey soil. However, the pores within a sandy soil tend to be larger, while the pores in a clayey soil are smaller. A soil with a more well-developed structure has greater total porosity, lower bulk density, and more large pores.

$$\text{Bulk Density: } \rho_b = \frac{\text{(mass of dry soil)}}{\text{(total soil volume)}} \qquad (2.1)$$

Soil *porosity* is the ratio of the void space (air volume plus water volume) to the total soil volume.

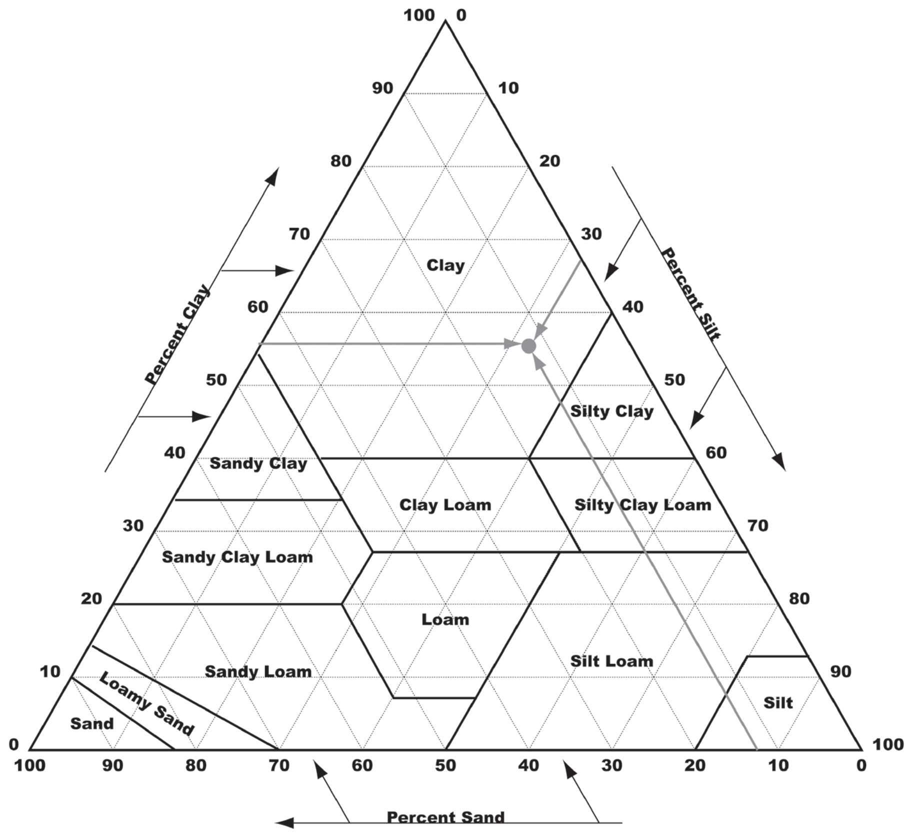

FIGURE 2.3. The soil textural triangle, illustrating the 12 soil textural classes.

$$\text{Porosity: } \phi = \frac{\text{(volume void space)}}{\text{(total soil volume)}} \tag{2.2}$$

$$\phi = \frac{(V_a + V_w)}{(V_a + V_w + V_p)} = \frac{(V_a + V_w)}{(V_t)}$$

where

V_a = the volume of air,

V_w = the volume of water,

V_p = the volume of soil particles, and

V_t = the total soil volume.

The internal architecture of the soil influences the water relations of the soil, including capillarity, infiltration, and percolation (discussed in greater detail later in this chapter). For example, soils with relatively high sand contents (sands, loamy sands, sandy loams) tend to have rapid infiltration and percolation rates, good aeration, and low water storage capacity. This is primarily due to the high proportion of larger pores and low surface area associated with the sandy soil (Fig. 2.4a). Conversely, finer textured soils tend to have slow rates of infiltration and percolation and poor aeration, mainly because of the lack of large pores that readily transmit water (Fig. 2.4b). However, finer textured soils that have well-developed structure, and therefore have more large pores created by the voids between individual peds, may also have rapid infiltration and percolation rates and good aeration (Fig. 2.4c). Clayey soils have a high water storage capacity, but many of the small pores hold water too tightly to be readily available to plants.

The minerals that make up the clay particles in the soil—mostly secondary aluminosilicates, are also more chemically active than the predominant minerals of the silt and sand particles, which are mostly silica. Clay particles have a much higher cation exchange capacity, which gives clayey soils a greater ability to retain plant nutrients. The high surface area and cation exchange capacity of clay particles also promotes interactions between clay and organic matter particles, which fosters greater organic matter retention in finer textured soils.

Along with soil texture, the other prominent property that greatly influences soil properties and processes is soil acidity, as quantified by pH. Soil acidity mainly influences the solubility of various elements in the soil, particularly some plant nutrients that influence productivity of wetland environments. At low pH values (<5.8) the availability of certain plant nutrients—such as phosphorus, nitrogen, calcium, and magnesium—may be limited. Microbial activity is also diminished when soil acidity is high. Conversely, aluminum and manganese availability is increased and may reach levels toxic to some plants. At high pH values (>7.5) the availability of phosphorus, iron, manganese,

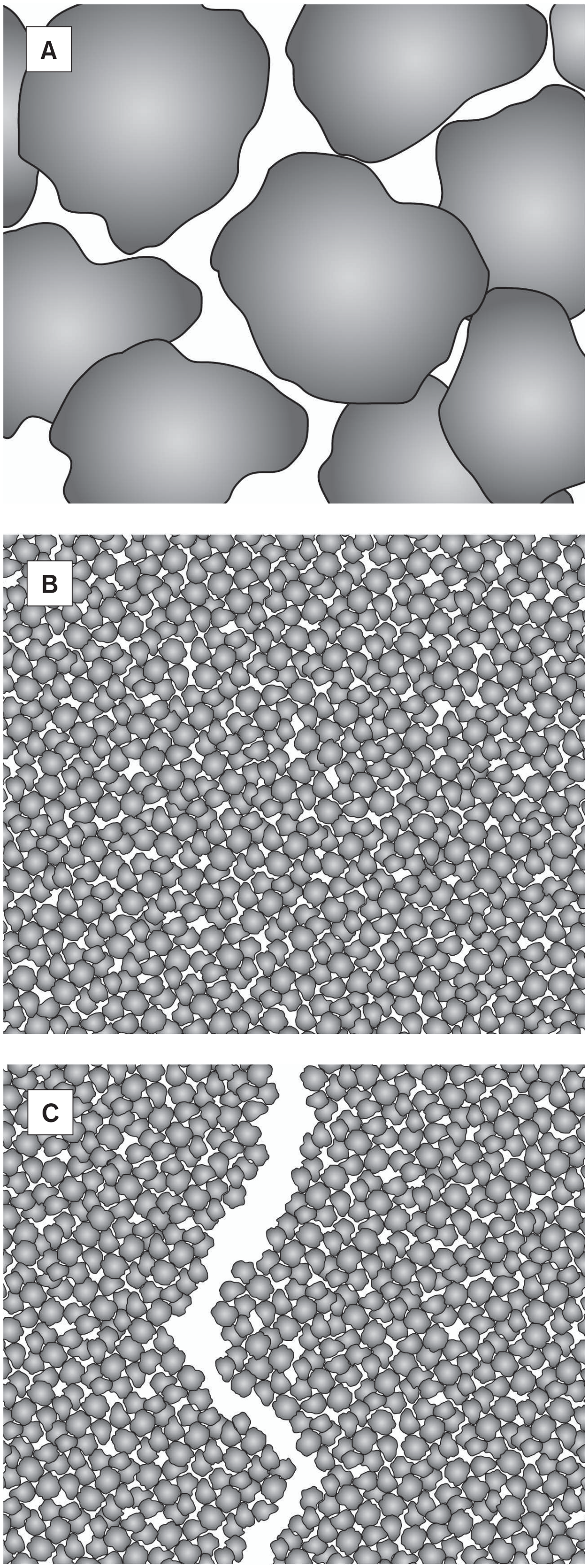

FIGURE 2.4. (A) Interparticle voids are relatively large between sand particles, creating numerous macropores. (B) These interparticle voids are much smaller between finer soil particles. (C) Structural development creates macropores between peds, which allows for greater water and air movement even in clayey soils.

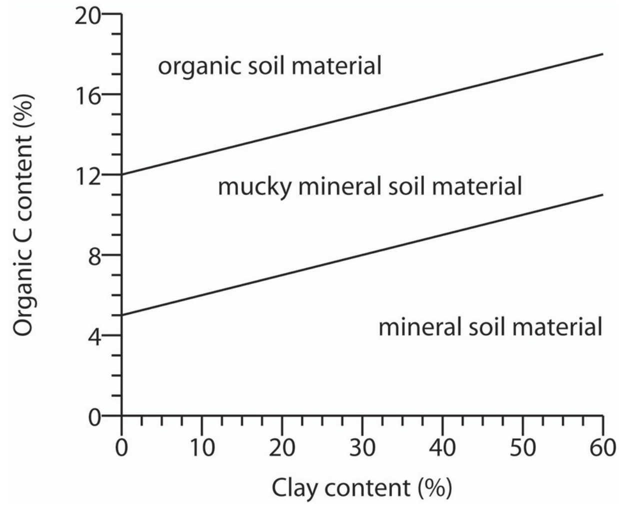

FIGURE 2.5. Organic soil materials are defined by the relationship between clay content and organic C content.

copper, and zinc is limited. The pH of a soil is controlled by the relative amounts of acidic (H+ and Al^{3+}) vs. nonacidic (Ca^{2+}, Mg^{2+}, K^+, Na^+) elements retained within the cation exchange capacity of a soil. If the soil parent materials are low in nonacidic cations, the resulting soil will also be low in nonacidic cations. Precipitation and organic matter decomposition add acidic cations to the soil, while groundwater influx tends to be a source of nonacidic cations. Wetlands receiving groundwater discharge are often less acidic than wetlands that receive most of their water through precipitation. Leaching and plant uptake remove nonacidic cations and concentrate acidic cations within the cation exchange complex.

The soil property most commonly associated with wetland soils is increased organic matter content. The prolonged saturated and anaerobic conditions in wetland soils slow organic matter decomposition and lead to organic matter accumulation. Organic matter, specifically humus, in a mineral soil promotes aggregation and structural stability, lowers bulk density, increases porosity, and leads to higher infiltration and percolation rates. Organic matter also contains significant amounts of plant nutrients (in unavailable forms), which can be converted to available forms during organic matter decomposition. The complex humus molecules also add to the cation exchange capacity of the soil.

If the organic C content is greater than 12% to 18%, depending on the clay content (Fig. 2.5), the soil material is considered organic. Soils dominated by organic soil materials have a low bulk density, high porosity, and a high water holding capacity; however, water movement through organic soil materials is generally slow. Although the nutrient content of the organic soil materials is high, much is not in available forms. The cation exchange capacity of organic soil materials is high, but the exchange complex is dominated by acidic cations, such that the pH of organic soil materials is generally low.

TABLE 2.1
The Six Master Horizons and the Distinguishing Characteristics of Each

Master Horizon	Description
O	Layer dominated by *organic material*
A	*Mineral* layer formed at the surface (or below an O or another A horizon) characterized by an *accumulation of humified organic matter,* or having properties resulting from cultivation or other agricultural activities
E	*Mineral* layer characterized by an *eluvial loss* of silicate clay, iron, and/or aluminum, leaving a concentration of sand- and silt-sized particles of quartz and/or other resistant minerals
B	*Mineral* layer dominated by one or more of the following: (1) *illuvial accumulation* of silicate clay, iron, aluminum, humus, carbonates, gypsum, and/or silica; (2) carbonate removal; (3) non-illuvial coatings or residual concentration of iron and/or aluminum sesquioxides; (4) structure development; (5) brittleness
C	*Mineral* layer that has been mostly *unaffected by pedogenic processes*
R	Hard *bedrock* layer

TABLE 2.2
Subordinate Soil Horizon Distinctions That Commonly Occur in Hydric Soils

Subordinate Distinction	Description
a	Highly decomposed organic material (sapric)
b	Buried genetic horizon in a mineral soil
e	Organic material of intermediate decomposition (hemic)
f	Frozen soil or water (permanent ice)
g	Strong gleying
h	Illuvial accumulation of organic matter
i	Slightly decomposed organic material (fibric)
p	Plowing or similar disturbance
s	Illuvial accumulation of sesquioxides and organic matter
t	Accumulation of silicate clay
w	Development of color or structure but with no illuvial accumulation
x	Fragic or fragipan characteristics

Soil Profiles

Soils properties differ with depth. Water movement over and through the soil (1) adds and removes materials, such as through erosion and deposition; (2) alters materials, such as through organic matter decomposition; and (3) redistributes materials within the profile, such as clay accumulation in the subsoil. These processes naturally lead to the development of layers within the soil. These layers are not depositional—they form in place as the soil develops from the parent material. The various soil horizons found with depth within a soil are approximately parallel to the soil surface. Each horizon will differ in its color, texture, structure, and/or other soil properties from the layers immediately below and above. These layers strongly affect the flow and distribution of water and the distribution of biological activity such as root growth, fungal and mycorrhizal growth (see Chapters 4 and 5), and animal burrowing and feeding (see Chapter 6).

Soil horizons are named in reference to their most important distinguishing characteristics. The master horizons, of which there are six, are designated by capital letters (Table 2.1). Lowercase letters sometimes follow the master horizon designations; these subordinate distinctions (Table 2.2) specify other important characteristics of the horizon. At the soil surface many wetland soils have an O horizon composed of vegetative detritus (leaves, needles, twigs, etc.) in various states of decay along with living vegetative matter. Wetland soils in warm humid climates with high biological activity will typically have thinner O horizons, while in cool and cold humid climates the O horizons tend to be thicker. The first mineral soil layer is the A horizon, which is often synonymous with the topsoil. A horizons can be of any texture, but are usually loamy or sandy relative to subsoil materials. A horizons are characterized by darker colors (low value, low chroma) because of their high organic matter content.

Depending on the characteristics of the parent material from which a soil profile was formed (as well as other soil-forming factors), there may be one or more B horizons below the A horizon. The B horizons are the subsoil and are normally characterized by the accumulation of materials translocated from upper portions of the soil profile. Common B horizon types are those that feature high contents of clay (Bt), organic material (Bh), or iron (Bs) that have been removed from the A horizon through eluviation by infil-

TABLE 2.3
Types of Soil Parent Materials, Their Characteristics, and Their Relationship to Soil Profile Properties

Soil Parent Material Type	General Description of Origin	Typical Geomorphic Position	Typical Soil Properties
Residuum	Weathered in place from underlying rock	Uplands—ridgetops and hillslopes	Highly variable; profiles typically include A, E, B, and C horizons
Colluvium	Weathered from upslope soil and bedrock carried by gravity; transport usually triggered by slope disturbance processes	Lower portions of hillslopes	Highly variable; profiles typically include A, E, B, and C horizons
Alluvium	Water borne materials such as river sediment deposits	Floodplains and stream terraces	Sandy or silty; no B horizon
Lacustrine	Material formed by sediment deposition on lake bottoms	Current and former lake beds	Silty or clayey
Aeolian	Wind-deposited material	Downwind of current and former desert environments	Sandy or silty
Glacial till	Material deposited in contact with a glacier	Anywhere affected by glaciation	Mixed sands, silts, and gravels; may be very dense with poor drainage
Glacial outwash	Material deposited by glacial meltwater	Broad glacial plains	Mixed sands and gravels; typically very porous; high conductivity
Marine	Material deposited by marine processes	Coastal plains	Mixed layers of sands and clays

trating water and concentrated by illuviation in the B horizon. Weakly developed B horizons (Bw) and gleyed B horizons (Bg) are also common. Below the solum, or combined A and B horizons, is often found the C horizon(s). These are composed of less-weathered parent material. The physical, chemical, and mineralogical characteristics of the parent materials can have a profound effect on the properties of overlying soil horizons (Table 2.3). Consolidated bedrock, when found within the soil profile, is labeled as an R layer. Between the A and B horizons, some soils feature a light-colored E horizon from which clay particles and organic matter have been eluviated.

Not every soil has all six master horizons. O horizons are not common, especially in disturbed or managed soils such as agricultural land, where any horizons within the plow layer are mixed to form an Ap horizon. E horizons are found mainly in more highly weathered soils in warmer and moister climates. Young soils, typified by alluvial floodplain soils, or soils formed from parent materials highly resistant to weathering often lack a B horizon, and feature an O or A horizon directly overlying C horizons.

Profile data from several seasonally saturated soils (Table 2.4) illustrate just a few of the variable horizon sequences that are commonly observed in and near wetland environments. Soils that experience prolonged saturation at or near the soil surface may develop thick O horizons, as is seen in the pocosin soil (Table 2.4). Below the highly decomposed (sapric) plant material (Oa horizons) there is little soil development in the mineral parent materials. These horizons are gleyed (Cg horizons) because of near-continuous saturated and anaerobic conditions. Wet mineral soils in some environments, such as in poorly drained soils in drainageways (Table 2.4), show an accumulation in organic matter to greater depths, forming several black (low value, low chroma) A horizons, with gleyed (Bg) horizons below. Distinct redox concentrations throughout these horizons are further evidence of prolonged saturated and anaerobic conditions. Upland soils can also have water tables at or near the surface, especially if they occur in depressional landscape positions (Table 2.4). Impermeable subsoil horizons, such as fragipans (Btx horizons), further contribute to the development of saturated and reducing conditions by promoting perched water tables. The Btg horizion immediately above the fragipan is evidence of the seasonally saturated conditions at a depth of 20 cm in this soil. The Btx horizons are not gleyed, but the macropores are lined with depleted (high value, low chroma) soil material.

Soil Processes

The shallow water tables and saturated soil conditions that are associated with wetland soils initiate a series of biogeo-

TABLE 2.4
Selected Morphologic Properties by Horizon of Four Seasonally Saturated Soils

Horizon	Depth (cm)	Matrix Color	Redox Concentrations[a]	Redox Depletions[a]
Pocosin				
Oa1	0–5	7.5YR 3/2		
Oa2	5–60	7.5YR 3/1		
Oe	60–85	2.5YR 3/2		
Oa3	85–202	5YR 3/2		
A	202–228	5Y 3/1		
Cg1	228–237	5Y 3/1		
Cg2	237–240	5Y 4/1		
Drainageway				
Ap	0–20	10YR 2/1	c f 7.5YR 4/6	
A2	20–53	N 2/0	f f 7.5YR 4/6	
A3	53–64	10YR 2/1	f f 7.5YR 4/6	
Bg1	64–88	2.5Y 4/2	m f 7.5YR 4/6	
Bg2	88–102	2.5Y 6/2	m f 7.5YR 4/7	
Bg3	102–155	2.5Y 6/2	m c 7.5YR 4/6	c m 5BG 5/1
Upland depression				
A	0–8	10YR 3/1		
E	8–20	10YR 6/2		
Btg	20–46	10YR 6/2	c m 7.5YR 5/8	c m 10YR 4/1
Btx1	46–71	10YR 4/3	m m 7.5YR 5/6	m m 10YR 6/2
Btx2	71–117	7.5YR 4/4	m m 7.5YR 5/6	m m 10YR 6/1
R	117+			
Terrace				
A	0–9	10YR 3/1		
Bw	9–22	10YR 5/3	c m 5YR 4/6	
C1	22–59	2.5Y 6/4	c m 5YR 4/6	c m 2.5Y 7/2
C2	59-85	2.5Y 6/4	c m 5YR 4/6	c m 5Y 7/2
Cg1	85-179	2.5Y 7/2	c m 7.5YR 6/4	
Cg2	179-210	2.5Y 7/1	f f 7.5YR 6/4	

[a] First letter indicates abundance (f = few, c = common, m = many); second letter(s) indicate size (f = fine, m = medium, c = coarse).

chemical processes that create the special ecological environment of wetland systems and control the functions and values of wetlands. These biogeochemical processes are primarily mediated by the microbial community, and that biological perspective is covered in detail in Chapter 4. Here we focus on the mineralogy and chemistry of biogeochemical reactions in wetland soils.

There is a general progression that occurs as a soil becomes saturated. As the water tables rise, air that is held in the soil pores is displaced by water (although a small fraction of pores retain entrapped air, so the degree of saturation never reaches 100%). The rate of oxygen diffusion into the soil is greatly diminished in a saturated soil. If temperature and bioavailable carbon are not limiting, microbes quickly deplete the oxygen that is trapped in the pores or dissolved in the soil solution of a saturated soil. Subsequently, the activity of facultative and obligate anaerobic microbes increases. These microbes function either as autotrophs, which may reduce Fe and Mn and employ the electron in ATP production, or as heterotrophs, which oxidize

TABLE 2.5
Order of Utilization of Electron Acceptors in Soils and Measured Potential of These Reactions in Soils

Reaction	Electrode Potential, pH 7 (v)	Measured Redox Potential in Soils (v)
$\frac{1}{2} O_2 + 2e^- + 2H^+ \leftrightarrow H_2O$	0.82	0.6 to 0.4
$NO_3^- + 2e^- + 2H^+ \leftrightarrow NO_2^- + H_2O$	0.54	0.5 to 0.2
$MnO_2 + 2e^- + 4H^+ \leftrightarrow Mn^{2+} + 2H_2O$	0.4	0.4 to 0.2
$FeOOH + e^- + 3H^+ \leftrightarrow Fe^{2+} + 2H_2O$	0.17	0.3 to 0.1
$SO_4^{2-} + 6e^- + 9H^+ \leftrightarrow HS^- + 4H_2O$	–0.16	0 to –0.15
$H^+ + e^- \leftrightarrow \frac{1}{2} H_2$	–0.41	–0.15 to –0.22
$(CH_2O)_n \leftrightarrow {}^{n}/_{2}\, CO_2 + {}^{n}/_{2}\, CH_4$	—	–0.15 to –0.22

SOURCE: After Bohn et al. (1985).

organic material and use Fe and Mn as electron acceptors during respiration.

OXIDATION-REDUCTION REACTIONS

In theory, the utilization of available oxidants dictates preferential use of the species that provides the greatest amount of energy to the microbes. In the soil system the order of electron acceptor preference is:

$$O_2 > NO_3^- > Mn(III\ or\ IV) > Fe(III) > SO_4^{2-} > H^+ > C.$$

The half-reactions that represent the reduction of each of these species are used to calculate the electrode potential associated with each reaction (Table 2.5). For a hypothetical reduction half reaction:

$$Ox + ne^- + mH+ \leftrightarrow Red + m/2\ H_2O$$

The electrode potential, *Eh,* is calculated as:

$$Eh = Eh^\circ - \frac{RT}{nF} \ln \frac{(Red)}{(Ox)\ (H^+)^m} \tag{2.3}$$

where Ox and Red are the oxidized and reduced species, respectively, Eh° is the standard electrode potential, R is the gas constant, T is the absolute temperature, F is the Faraday constant, and values in parentheses are activities.

Redox potentials at which reduction of O_2, NO_3^-, Mn(III or IV), Fe(III), SO_4^{2-}, and H^+ occur in the soil are not as discrete as the calculated electrode potentials (Table 2.5), with significant overlap among the observed ranges. This occurs because of the nature of redox potential and its measurement: (1) the calculated electrode potential is an equilibrium potential, but the soil system does not reach oxidation-reduction equilibrium because of the constant additions and losses of oxidants and reductants within the system (Bohn et al. 1985); (2) the potential that is measured by the platinum electrode represents multiple oxidation-reduction reactions occurring in the soil at the electrode surface; and (3) each reaction is a function of the concentration of reactants and the activity of selective microbes that facilitate oxidation and reduction reactions around the electrode. Therefore, electrode potentials and redox potentials are not equivalent.

Certain microbes catalyze the reduction of Fe(III) and Mn(III or IV) oxides, hydroxides, and oxyhydroxides (collectively *oxides*). When these microbes, such as *Micrococcus lactilyticus* and *Thiobacillus thiooxidans* (Zajic 1969), contact Fe and Mn oxides on soil particle surfaces, they reduce the Fe(III) or Mn(III or IV) to Fe(II) and Mn(II). The more soluble Fe(II) and Mn(II) ions readily dissolve into the soil solution (Fischer 1988). Depending on hydraulic and chemical gradients in the soil solution, the Fe^{2+} or Mn^{2+} may: (1) remain in the vicinity of the original soil particle surface until oxidizing conditions return; (2) become adsorbed to the cation exchange sites in the soil; (3) be translocated locally until an oxidizing environment is encountered and is reprecipitated as a Fe or Mn oxide mineral; or (4) be leached from the soil system. Depending on the fate of the reduced Fe or Mn, various morphological features may develop, such as low-chroma mottles in a high-chroma matrix, high-chroma mottles in a low-chroma matrix, or a gleyed soil.

Soil saturation and development of anoxic conditions causes (Ponnamperuma 1972): (1) a decrease in redox potential; (2) neutralization of pH; (3) changes in specific conductance and ion strength; (4) changes in certain mineral

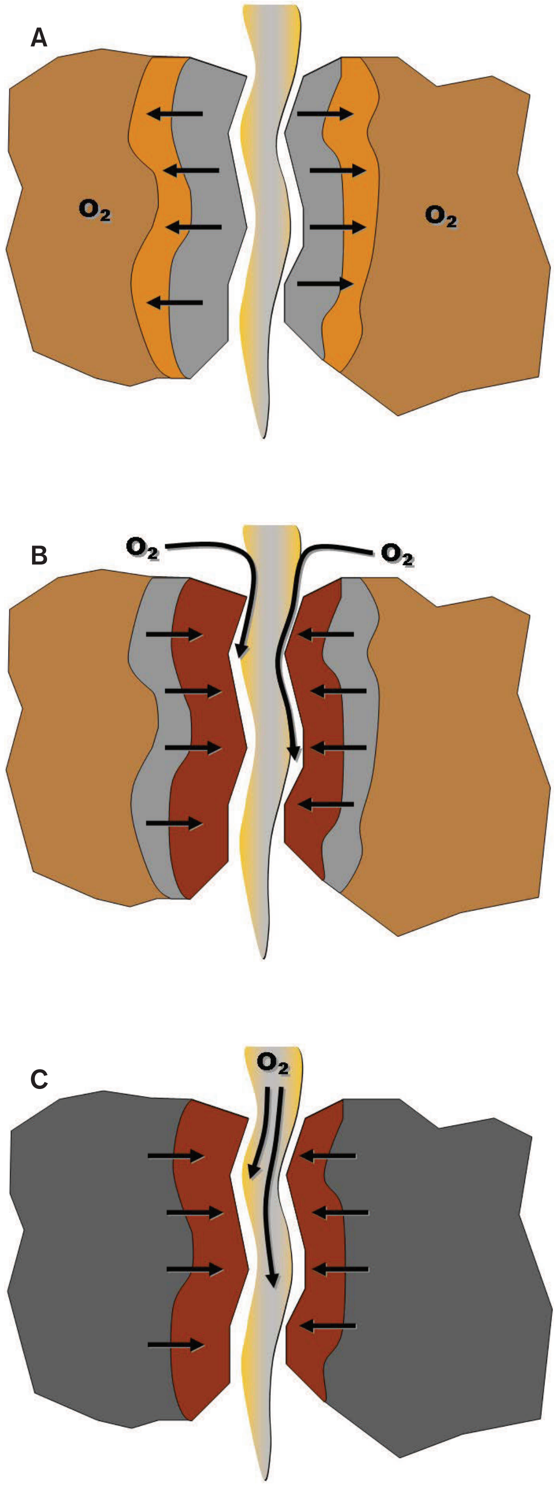

FIGURE 2.6. Models of redoximorphic feature formation. (A) Saturated and reduced pore and adjacent soil is the site of Fe(III) reduction and an aerated and oxidized matrix is the site of Fe(II) oxidation. (B) Saturated and reduced matrix is the site of Fe(III) reduction and an aerated and oxidized pore and adjacent soil is the site of Fe(II) oxidation. (C) Saturated and reduced matrix is the site of Fe(III) reduction and an oxidized rhizosphere is the site of Fe(II) oxidation.

equilibria; (5) ion exchange reactions; and (6) sorption and desorption of ions. In a mixed system, such as the soil, the dominant redox couple determines the redox potential (Ponnamperuma 1972). The order of oxidant utilization and associated potential of these reactions in soils (Table 2.5) indicates the redox potential that may be expected when that reaction is controlling the redox chemistry of a soil.

REDOXIMORPHIC FEATURE FORMATION

There are several theories explaining the formation of redoximorphic features under different hydrologic regimes (Veneman et al. 1976, Fanning and Fanning 1989, Vepraskas 1992). The location of saturated and aerated soil zones, and therefore the source of Fe(III) reduction within the soil relative to pores or the soil matrix, distinguishes between the hypothesized mechanisms of redoximorphic feature formation. Models of redoximorphic feature formation can be divided into two basic categories (Fig. 2.6): (1) a saturated and reduced pore and adjacent soil is the site of Fe(III) reduction and an aerated and oxidized matrix is the site of Fe(II) oxidation (Fig. 2.6a) or (2) a saturated and reduced matrix is the site of Fe(III) reduction and an aerated and oxidized pore and adjacent soil is the site of Fe(II) oxidation (Fig. 2.6b). Both of these types of redoximorphic features are readily observed in seasonally saturated soils (Table 2.4). The redox depletions in the third Bg horizon of the drainageway soil and the Btx horizons of the upland depression soil (Table 2.4) formed when the macropores between the peds were strongly reducing and Fe was translocated away from the pore (Fig. 2.6a). Most of the redox concentrations in the subsoil horizons of the drainageway, upland depression, and terrace soils (Table 2.4) were formed when oxygen was reintroduced via the macropores and the reduced Fe was reoxidized along the pore (Fig. 2.6b). A special case of this second mechanism of redoximorphic feature formation is seen prominently in dark A horizon materials, such as the upper horizons of the drainageway soil (Table 2.4). Some wetland plants are able to transport O_2 down to their roots. This can create an oxidized rhizosphere in which reduced Fe from the surrounding saturated soil will oxidize and precipitate around the root (Fig. 2.6c). This is often the only type of redoximorphic feature seen in surface horizons of wetland soils with thick, dark A horizons.

ORGANIC MATTER DECOMPOSITION AND ACCUMULATION

Organic matter is an important component of all wetland systems because it is the energy source for the microbial activity that drives the development of anaerobic and reducing conditions (see Chapter 4). The subsequent soil biogeochemical processes often lead to the accumulation of greater amounts of organic soil matter that, along with the presence of Fe-based redoximorphic features, is the property most commonly associated with wetland soils.

Soil microorganisms (bacteria and fungi) play the most significant role in organic matter decomposition in soils. In well-drained, aerobic soils, the rate of organic matter decomposition is often much greater than the rate of organic matter deposition from above- and below-ground biomass (leaves, stems, roots, macroorganisms, microorganisms). As a result, the equilibrium level of soil organic matter can be quite low (e.g., <2%). However, under anaerobic conditions that develop in saturated wetland soils, the aerobic decomposers no longer function and the facultative and obligate anaerobic microorganisms are left to decompose organic matter. These organisms do not derive as much energy when electron acceptors other than O_2 are used (Table 2.5), and organic matter decomposition can occur at a much slower rate in saturated and anaerobic soils (see Chapter 4). Consequently, organic matter inputs can be much greater than outputs and the equilibrium level of soil organic matter is higher in wetland soils (see pocosin and drainageway soils in Table 2.4).

Although organic matter accumulation is typical of wetland soils, not all wetland soils have accumulated enough organic matter to have an organic soil horizon at the soil surface. The presence of an O horizon or a black A horizon at the soil surface are commonly associated with wetland mineral soils.

BOX 2.1. THE MANDATORY TECHNICAL CRITERIA FOR HYDRIC SOILS

1. All Histels except Folistels, and Histosols except Folists; or
2. Soils in Aquic suborders, great groups, or subgroups, Albolls suborder, Historthels great group, Histoturbels great group, Pachic subgroups, or Cumulic subgroups that are:
   a. Somewhat poorly drained with a water table[a] equal to 0.0 foot (ft) from the surface during the growing season; or
   b. Poorly drained or very poorly drained and have either:
      i. a water table equal to 0.0 ft during the growing season[b] if textures are coarse sand, sand, or fine sand in all layers within 20 inches (in), or for other soils; or
      ii. a water table at less than or equal to 0.5 ft from the surface during the growing season if permeability is equal to or greater than 6.0 in/hour (h) in all layers within 20 in; or
      iii. a water table at less than or equal to 1.0 ft from the surface during the growing season if permeability is less than 6.0 in/h in any layer within 20 in; or
3. Soils that are frequently[c] ponded for long duration[d] or very long duration[e] during the growing season; or
4. Soils that are frequently flooded for long duration or very long duration during the growing season.

SOURCE: www.nrcs.usda.gov/wps/portal/nrcs/main/soils/use/hydric/, Soil Survey Staff 2010.

[a] Water table = the upper surface of ground water where the water is at atmospheric pressure. In the Map Unit Interpretation Record (MUIR) database, entries are made for the zone of saturation at the highest average depth during the wettest season. It is at least six inches thick and persists in the soil for more than a few weeks. In other databases, saturation, as defined in *Soil Taxonomy* (Soil Survey Staff, 1999), is used to identify conditions that refer to water table in Criteria 2.

[b] Growing season = the portion of the year when soil temperatures are above biologic zero at 50 cm; defined by the soil temperature regime.

[c] Frequently = flooding, ponding, or saturation is likely to occur often under usual weather conditions (more than 50 percent chance in any year, or more than 50 times in 100 years).

[d] Long duration = a single event lasting 7 to 30 days.

[e] Very long duration = a single event lasting longer than 30 days.

Legal Differentiation of Wetland Soils

In the United States, hydric soils are identifying characteristics, along with hydrophytic vegetation and wetland hydrology, that legally define wetlands (see Chapter 8). Hydric soils are specifically defined as soils that formed under conditions of saturation, flooding, or ponding long enough during the growing season to develop anaerobic conditions in the upper part (Federal Register 1994). From this definition, the USDA Natural Resources Conservation Service (NRCS 2010) developed a set of mandatory technical criteria for hydric soils. These criteria (Box 2.1) serve mainly as a way to retrieve a list of likely hydric soils from a database of soil information; however, the criteria can also be used as indicators for identifying hydric soils in the field. Hydric soil lists are developed and updated using these criteria and can be used in conjunction with published soil survey reports to generate preliminary inventories of hydric soils in an area. It is important to note that on-site field verification of the presence of hydric soils is required because soil survey maps cannot represent all soils within an area, only soil bodies large enough to be delineated at the scale of the map (usually larger than 1.2 ha). Also, placement on a hydric soil list does not guarantee that a soil is indeed hydric. It only indicates that the range in properties associated with a given soil in a map unit overlap with those of the technical criteria.

Most hydric soil determinations are based on field indicators. The 1987 *USACE Wetlands Delineation Manual* lists a series of field indicators intended to be used as general guidelines for field identification of hydric soil (Box 2.2). More detailed and specific field indicators (NRCS 2002) have been developed for on-site identification and delineation of hydric soils. These indicators (Box 2.3) are observable soil morphological properties that form when the soil

BOX 2.2. FIELD INDICATORS OF HYDRIC SOILS

1. Organic soils[a]
2. Histic epipedons[a]
3. Sulfidic material[b]
4. Aquic or peraquic moisture regime[a]
5. Direct observation of reducing soil conditions with α-α dipyridyl indicator solution
6. Gleyed, low chroma and low chroma/mottled soils
   a. Gleyed soils
   b. Low chroma soils and mottled soils
7. Iron and manganese concretions

SOURCE: According to the 1987 *USACE Wetlands Delineation Manual.*

[a] As defined in *Keys to Soil Taxonomy* (Soil Survey Staff, 2010).

[b] As evidenced by hydrogen sulfide, or rotten egg odor.

is saturated, flooded, or ponded long enough during the growing season to develop anaerobic conditions in the upper part. Some indicators can be applied to all soil types, while others can be applied only to sandy soils or only to loamy and clayey soils. The variety of soil morphologies by which hydric soil conditions can be expressed is evidenced by the length of this list of indicators. However, the indicators are regionally specific, so not all of them are applicable in all places. Normally, within a region, there are a small number of indicators that can reasonably be expected to be used in most circumstances.

Use of the indicators is comparative. After exposing and describing a soil profile to a depth of at least 50 cm, the descriptions of the field indicators are then compared to the field description. For example, the thick organic layers of the pocosin soil (Table 2.4) more than adequately meet the requirements of indicator A1, which requires a minimum of 40 cm of organic soil material in the upper 80 cm of soil. A thinner (20–40 cm) accumulation of organic soil materials at the surface might meet the requirements of indicator A2 or A3. Even less organic soil material at the surface may express indicator A7, A8, A9, or A10. The drainageway soil (Table 2.4) also has an accumulation of organic matter, but not organic soil materials. Below the thick, dark A horizons is a layer with a depleted matrix. For this loamy soil, indicator F6 applies. If the surface horizon were thinner, indicators F3 or A11 might apply. If the surface horizon had hue N like the second A horizon, the requirements of indicator A12 would be met. For soils without organic soil materials or thick, dark surfaces, it is the subsoil color that most often is the reliable indicator of seasonally saturated and reducing conditions. Specifically, the presence of gleyed matrix colors or the presence of a depleted (high value, low chroma) matrix is often used to identify hydric soils. Depending on the exact Munsell value and chroma, the presence of redoximorphic features may be required along with a depleted matrix. For the upland depression soil (Table 2.4), the Btg horizon, which starts at a depth of 20 cm, has a depleted matrix and meets indicator F3. This horizon has redox concentrations, but the relatively high value means that redox concentrations are not required to meet this indicator. Conversely, when examining the terrace soil (Table 2.4), the presence of redox concentrations starting at a depth of 9 cm is not enough to meet any hydric soil indicator. This soil does experience high water table conditions during the year, as evidenced by the high value and low chroma colors deeper in the profile, but prolonged saturated and reducing conditions do not occur close enough to the surface to meet the definition of a hydric soil.

Hillslope and Wetland Hydrology

The hydroperiod and hydropattern of a wetland are among the dominant controls on soils, plant and animal communities, primary productivity, and decomposition within wetlands. *Hydroperiod* is a statistically ill-defined term that refers to the general seasonal pattern of surface inundation depth. Nuttle (1997, p. 82) says "the pattern of water-level fluctuations in a wetland is its hydroperiod," although this simple definition works better for the alternative term *hydropattern,* which refers to the typical water-level time series of a wetland. Several researchers, working mostly on Everglades restoration, have suggested that *hydropattern* would be a better term for the hydrologic dynamics of a wetland and that *hydroperiod* should refer only to the dates and duration of surface inundation (e.g., Acosta and Perry 2001, 2002; King et al. 2004). While *hydroperiod,* used in its broadest sense, is still the dominant term in wetland literature, this chapter will use *hydropattern* to discuss the time series of water-level variation from this point forward. Classification or description of wetlands by hydropattern has become a common framework from which to explore and explain wetland ecology. Understanding a wetland's hydropattern, and the variation of hydopatterns among wetlands, requires a general understanding of the hydrologic processes that deliver and export water into and out of a wetland.

Hillslope Hydrologic Processes

The hydrologic pathways that define a wetland's water budget and hydropattern depend on the landscape position of a wetland. For instance, a wetland may occur over a low conductivity layer on a ridge and feature simple, largely vertical, hydrologic fluxes including precipitation, evapotranspiration, and percolation through the impeding layer. Alternatively, a wetland may sit at the edge of a floodplain at the base of the slope and receive precipitation, shallow groundwater from the adjacent hillslope, deep regional groundwater, and occasional river overflows while also exporting water as evapotranspiration, leakage

BOX 2.3. FIELD INDICATORS OF HYDRIC SOILS IN THE UNITED STATES

ALL SOILS

A1 Histosol or Histel[a]—Soil classifies as a Histosol (except Folist) or as a Histel (except Folistel).

A2 Histic Epipedon[a]—Soil has a histic epipedon underlain by mineral soil material with a chroma of 2 or less.

A3 Black Histic—Soil has a layer of peat, mucky peat, or muck 20 cm (8 in) or more thick that starts within the upper 15 cm (6 in) of the soil surface; has a hue of 10YR or yellower, a value of 3 or less, and a chroma of 1 or less; it is underlain by mineral soil material with a chroma of 2 or less.

A4 Hydrogen Sulfide—Soil has hydrogen sulfide odor within 30 cm (12 in) of the soil surface.

A5 Stratified Layers—Soil has several stratified layers starting within the upper 15 cm (6 in) of the soil surface. At least one of the layers has a value of 3 or less with a chroma of 1 or less, or it is muck, mucky peat, peat, or mucky modified mineral texture. The remaining layers have a chroma of 2 or less. For any sandy material that constitutes the layer with a value of 3 or less and a chroma of 1 or less, at least 70% of the visible soil particles must be masked with organic material when viewed through a 10× or 15× hand lens. Observed without a hand lens, the particles appear to be close to 100% masked.

A6 Organic Bodies—Soil has 2% or more organic bodies of muck or a mucky modified mineral texture starting within 15 cm (6 in) of the soil surface.

A7 5 cm Mucky Mineral—Soil has a layer of mucky modified mineral soil material 5 cm (2 in) or more thick, starting within 15 cm (6 in) of the soil surface.

A8 Muck Presence—Soil has a layer of muck with a value of 3 or less and a chroma of 1 or less within 15 cm (6 in) of the soil surface.

A9 1 cm Muck—Soil has a layer of muck 1 cm (0.5 in) or more thick with a value of 3 or less and a chroma of 1 or less starting within 15 cm (6 in) of the soil surface.

A10 2 cm Muck—Soil has a layer of muck 2 cm (0.75 in) or more thick with a value of 3 or less and a chroma of 1 or less starting within 15 cm (6 in) of the soil surface.

A11 Depleted Below Dark Surface[b,c]—Soil has a layer with a depleted or gleyed matrix that has 60% or more chroma of 2 or less, starting within 30 cm (12 in) of the soil surface, and having a minimum thickness of either (a) 15 cm (6 in), or (b) 5 cm (2 in) if the 5 cm consists of fragmental soil material. Loamy or clayey layer(s) above the depleted or gleyed matrix must have a value of 3 or less and a chroma of 2 or less. Any sandy material above the depleted or gleyed matrix must have a value of 3 or less and a chroma of 1 or less, and, viewed through a10× or 15× hand lens, at least 70% of the visible soil particles must be masked with organic material. Observed without a hand lens, the particles appear to be close to 100% masked.

A12 Thick Dark Surface[b,c]—Soil has a layer at least 15 cm (6 in) thick with a depleted or gleyed matrix that has 60% or more chroma of 2 or less starting below 30 cm (12 in) of the surface. The layer(s) above the depleted or gleyed matrix must have a value of 2.5 or less and a chroma of 1 or less to a depth of at least 30 cm (12 inches), and a value 3 or less and a chroma 1 or less in any remaining layers above the depleted or gleyed matrix. In any sandy material above the depleted or gleyed matrix, at least 70% of the visible soil particles must be masked with organic material when viewed through a 10× or 15× hand lens. Observed without a hand lens, the particles appear to be close to 100% masked.

A13 Alaska Gleyed—Soil has a mineral layer with a dominant hue of N, 10Y, 5GY, 10GY, 5G, 10G, 5BG, 10BG, 5B, 10B, or 5PB and with a value of 4 or more in more than 50% of the matrix. The layer starts within 30 cm (12 in) of the mineral surface and is underlain within 1.5 m (60 in) by soil material with a hue of 5Y or redder in the same type of parent material.

A14 Alaska Redox[b]—Soil has mineral layer that has a dominant hue of 5Y with a chroma of 3 or less, or a gleyed matrix, with 10% or more distinct or prominent redox concentrations occurring as pore linings with value and chroma of 4 or more. The layer occurs within 30 cm (12 in) of the soil surface.

A15 Alaska Gleyed Pores—Soil has a mineral layer that has 10% or more with a hue of N, 10Y, 5GY, 10GY, 5G, 10G, 5BG, 10BG, 5B, 10B, or 5PB with a value of 4 or more along root channels or other pores and that starts within 30 cm (12 in) of the soil surface. The matrix has a dominant hue of 5Y or redder.

A16 Coast Prairie Redox—Soil has a layer starting within 15 cm (6 in) of the soil surface that is at least 10 cm (4 in) thick and has a matrix chroma of 3 or less with 2% or more distinct or prominent redox concentrations occurring as soft masses and/or pore linings.

SANDY SOILS

S1 Sandy Mucky Mineral—Soil has a layer of mucky modified sandy soil material 5 cm (2 in) or more thick starting within 15 cm (6 in) of the soil surface.

(continued)

BOX 2.3 *(continued)*

S2 2.5 cm Mucky Peat or Peat—Soil has a layer of mucky peat or peat 2.5 cm (1 in) or more thick with a value of 4 or less and a chroma of 3 or less starting within 15 cm (6 in) of the soil surface, and underlain by sandy soil material.

S3 5 cm Mucky Peat or Peat—Soil has a layer of mucky peat or peat 5 cm (2 in) or more thick with a value of 3 or less and a chroma of 2 or less starting within 15 cm (6 in) of the soil surface, and underlain by sandy soil material.

S4 Sandy Gleyed Matrix[b]—Soil has a gleyed matrix that occupies 60% or more of a layer starting within 15 cm (6 in) of the soil surface.

S5 Sandy Redox—Soil has a layer starting within 15 cm (6 in) of the soil surface that is at least 10 cm (4 in) thick and has a matrix with 60% or more chroma of 2 or less with 2% or more distinct or prominent redox concentrations as soft masses and/or pore linings.

S6 Stripped Matrix—Soil has a layer starting within 15 cm (6 in) of the soil surface in which iron/manganese oxides and/or organic matter have been stripped from the matrix and the primary base color of soil material has been exposed. The stripped areas and translocated oxides and/or organic matter form a faintly contrasting pattern of two or more colors with diffuse boundaries. The stripped zones are 10% or more of the volume and are rounded.

S7 Dark Surface—Soil has a layer 10 cm (4 in) or more thick starting within the upper 15 cm (6 in) of the soil surface with a matrix value of 3 or less and a chroma of 1 or less. At least 70% of the visible soil particles must be masked with organic material when viewed through a 10× or 15× hand lens. Observed without a hand lens, the particles appear to be close to 100% masked. The matrix color of the layer directly below the dark layer must have the same colors as those described above or any color that has a chroma of 2 or less.

S8 Polyvalue Below Surface—Soil has a layer with a value of 3 or less and a chroma of 1 or less starting within 15 cm (6 in) of the soil surface. At least 70% of the visible soil particles must be masked with organic material when viewed through a 10× or 15× hand lens. Observed without a hand lens, the particles appear to be close to 100% masked. Directly below this layer, 5% or more of the soil volume has a value of 3 or less and a chroma of 1 or less, and the remainder of the soil volume has a value of 4 or more and a chroma of 1 or less to a depth of 30 cm (12 in) or to the spodic horizon, whichever is less.

S9 Thin Dark Surface—Soil has a layer 5 cm (2 in) or more thick, within the upper 15 cm (6 in) of the soil, with a value of 3 or less and a chroma of 1 or less. At least 70% of the visible soil particles in this layer must be masked with organic material when viewed through a 10× or 15× hand lens. Observed without a hand lens, the particles appear to be close to 100% masked. This layer is underlain by a layer or layers with a value of 4 or less and a chroma of 1 or less to a depth of 30 cm (12 in) or to the spodic horizon, whichever is less.

LOAMY AND CLAYEY SOILS

F1 Loamy Mucky Mineral—Soil has a layer of mucky modified loamy or clayey soil material 10 cm (4 in) or more thick starting within 15 cm (6 in) of the soil surface.

F2 Loamy Gleyed Matrix[b]—Soil has a gleyed matrix that occupies 60% or more of a layer starting within 30 cm (12 in) of the soil surface.

F3 Depleted Matrix[c]—Soil has a layer with a depleted matrix that has 60% or more chroma of 2 or less that has a minimum thickness of either (a) 5 cm (2 in) if 5 cm (2 in) is entirely within the upper 15 cm (6 in) of the soil, or (b) 15 cm (6 in) and starts within 25 cm (10 in) of the soil surface.

F6 Redox Dark Surface—Soil has a layer at least 10 cm (4 in) thick, is entirely within the upper 30 cm (12 in) of the mineral soil, and has either (a) a matrix value of 3 or less and a chroma of 1 or less and 2% or more distinct or prominent redox concentrations as soft masses or pore linings, or (b) a matrix value of 3 or less and a chroma of 2 or less and 5% or more distinct or prominent redox concentrations as soft masses or pore linings.

F7 Depleted Dark Surface—Soil has redox depletions with a value of 5 or more and a chroma of 2 or less in a layer at least 10 cm (4 in) thick, is entirely within the upper 30 cm (12 in) of the mineral soil, and has either (a) a matrix value of 3 or less and a chroma of 1 or less and 10% or more redox depletions, or (b) a matrix value of 3 or less and a chroma of 2 or less and 20% or more redox depletions.

F8 Redox Depressions—Soil is in closed depression subject to ponding with 5% or more distinct or prominent redox concentrations as soft masses or pore linings in a layer 5 cm (2 in) or more thick entirely within the upper 15 cm (6 in) of the soil surface.

F9 Vernal Pools—Soil is in closed depressions subject to ponding with a depleted matrix with 60% or more chroma of 2 or less in a layer 5 cm (2 in) thick entirely within the upper 15 cm (6 in) of the soil surface.

F10 Marl—Soil has a layer of marl that has a value of 5 or more starting within 10 cm (4 in) of the soil surface.

F11 Depleted Ochric—Soil has a layer 10 cm (4 in) or more thick that has 60% or more of the matrix with a value of 4 or more and a chroma of 1 or less. The layer is entirely within the upper 25 cm (10 in) of the soil surface.

F12 Iron/Manganese Masses—Soil is on flood plains with a layer 10 cm (4 in) or more thick with 40% or more chroma of 2 or less and 2% or more distinct or prominent redox concentrations as soft iron/manganese masses with diffuse boundaries. The layer occurs entirely within 30 cm (12 in) of the soil surface. Iron/manganese masses have a value of 3 or less and a chroma of 3 or less. Most commonly they are black. The thickness requirement is waived if the layer is the mineral surface layer.

F13 Umbric Surface—Soil is in depressions or other concave landforms with a layer 25 cm (10 in) or more thick, starting within 15 cm (6 in) of the soil surface, in which the upper 15 cm (6 in) has a value of 3 or less and a chroma of 1 or less and in which the lower 10 cm (4 in) has the same colors as above or any other color that has a chroma of 2 or less.

F16 High Plains Depressions—Soil is in closed depressions subject to ponding, with a mineral soil that has a chroma of 1 or less to a depth of at least 35 cm (13.5 in) and a layer at least 10 cm (4 in) thick within the upper 35 cm (13.5 in) of the mineral soil that has either (a) 1% or more redox concentrations as nodules or concretions, or (b) redox concentrations as nodules or concretions with distinct or prominent corona.

F17 Delta Ochric—Soil has a layer 10 cm (4 in) or more thick in which 60% or more of the matrix has a value of 4 or more and a chroma of 2 or less and there are no redox concentrations. This layer occurs entirely within the upper 30 cm (12 in) of the soil.

F18 Reduced Vertic[a]—Soil classifies as a Vertisols or Vertic intergrade and has a positive reaction to alpha-alpha-dipyridyl? that (a) is the dominant (60% or more) condition of a layer at least 10 cm (4 in) thick within the upper 30 cm (12 in), or at least 5cm (2 in) thick within the upper 15 cm (6 in), of the mineral or muck soil surface; (b) occurs for at least 7 continuous days and 28 cumulative days; and (c) occurs during a normal or drier season and month (within 16 to 84% of probable precipitation).

F19 Piedmont Flood Plain Soils—Soil is on an active floodplain and has a mineral layer at least 15 cm (6 in) thick, starting within 25 cm (10 in) of the soil surface, with a matrix (60% or more of the volume) chroma of less than 4 and 20% or more distinct or prominent redox concentrations occurring as soft masses or pore linings.

F20 Anomalous Bright Loamy Soils—Soil is located within 200 m (656 ft) of estuarine marshes or water and within 1 m (3.28 ft) of mean high water, a mineral layer at least 10 cm (4 in) thick, starting within 20 cm (8 in) of the soil surface, with a matrix (60% or more of the volume) chroma of less than 5 and 10% or more distinct or prominent redox concentrations occurring as soft masses or pore linings and/or depletions.

SOURCE: USDA-NRCS, 2010.

[a] As defined in *Keys to Soil Taxonomy* (Soil Survey Staff, 2010).

[b] Soils that have a gleyed matrix have the following combinations of hue, value, and chroma and the soils are not glauconitic:
(a) 10Y, 5GY, 10GY, 10G, 5BG, 10BG, 5B, 10B, or 5PB with a value of 4 or more and a chroma of 1; or
(b) 5G with a value of 4 or more and a chroma of 1 or 2; or
(c) N with a value of 4 or more; or
(d) (for testing only) 5Y, a value of 4 or more, and a chroma of 1.

[c] The following combinations of value and chroma identify a depleted matrix:
(a) Matrix value of 5 or more and a chroma of 1 with or without redox concentrations as soft masses and/or pore linings, or
(b) Matrix value of 6 or more and a chroma of 2 or 1 with or without redox concentrations as soft masses and/or pore linings, or
(c) Matrix value of 4 or 5 and a chroma of 2, and has 2% or more distinct or prominent redox concentrations as soft masses and/or pore linings, or
(d) Matrix value of 4 and a chroma of ,1 and has 2% or more distinct or prominent redox concentrations as soft masses and/or pore linings.

to the floodplain, and drainage to the river during flood recession. If the hillslope draining to this more complicated wetland were converted to row-crop agriculture or to suburban development, the wetland would also receive surface runoff. So, to understand a wetland's hydrology, one must understand basic hydrologic processes as well as how land use activities alter these processes.

Hillslope hydrology is a sub-branch of hydrology devoted to describing and explaining how water moves through the terrestrial landscape into surface waters. The effects of land management activities—such as cropping, silviculture, forest conversion (clearing and stump removal), and development—on downstream surface waters are best understood through the prism of hillslope hydrology. Furthermore, a wetland's water budget and its resulting hydropattern are best understood by framing a wetland with respect to the dominant hydrologic processes in the contributing watershed. Hillslope hydrology encompasses interception, infiltration, Horton overland flow, evapotranspiration, interflow, percolation, groundwater flow, and variable source area runoff (Fig. 2.7). Here we will explain each of these processes and how they affect the routing of water to and from wetlands.

Interception

Except in barren and denuded environments, precipitation first strikes vegetative surfaces (leaves, branches, stems, trunks, and vegetative detritus) before hitting the soil. Depending on the humidity within the canopy, a cer-

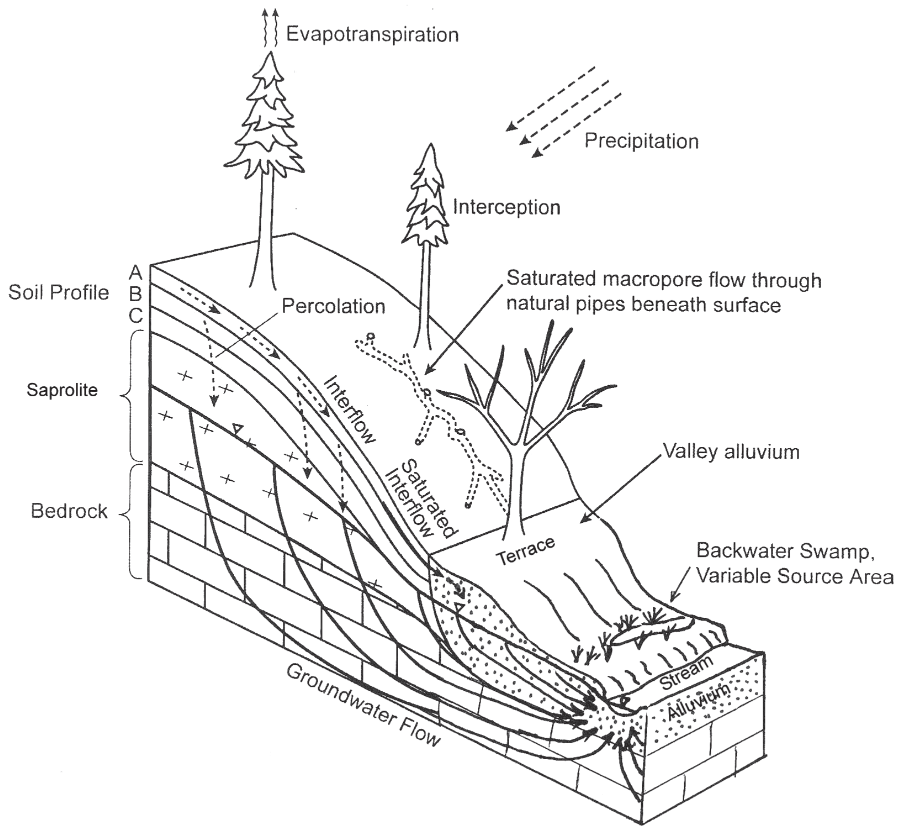

FIGURE 2.7. Hillslope flowpaths and hydrologic processes in a typical humid forest. Adapted from Atkinson (1998).

tain amount of the water temporarily sorbed onto vegetative surfaces evaporates before dripping to the forest floor. This process, well known to all children who play outdoors and use trees for temporary cover, is called *interception*. The amount of water a canopy can intercept is a function of *leaf area index*, or the ratio of vegetative surface area to the underlying ground surface area. Leaf area indices range from as little as 1.0 for short grasses and desert scrub, around 3.0–4.0 for grasslands and savannahs, 5.0–8.0 for temperate deciduous forests, and 8.0–18.0 for conifer forests and rainforests (Reichle 1981; Perry 1994; Barbour et al. 1998). In deciduous forests, interception varies seasonally with the changing leaf cover, and in all forests, leaf area index and interception vary with stand age and condition. On an annual basis, interception is greater in coniferous forests than in deciduous forests in the same climatic region.

The quantity of precipitation intercepted during individual storms is small—in the range of 0.7 to 2.0 mm (Shuttleworth 1992). However, since humid climatic regions often experience 50–90 precipitation events per year, annual interception totals comprise a significant fraction of annual precipitation (Table 2.6).

Human alterations of the landscape, through urbanization, agriculture, and forestry, have direct impacts on interception. By removing the overhead canopy and reducing leaf area indices, timber harvest reduces interception and increases the amount of precipitation reaching the soil. Along with reduced evapotransiration (discussed below), this usually contributes to elevated water tables and higher baseflow for several years following forest harvest.

There are rare geographic exceptions to the relationship between leaf area indices and interception. In cloud forests, or areas where dense fog is common, fog-drip occurs in a process that could be considered negative interception. In such areas, tree canopies act as mist-nets that condense cloud water. This condensation is perceived as fog-drip, and it results in a net increase in precipitation. Removal of the canopy in fog-drip areas leads to lowered water tables and reduced stream flows (Harr 1982).

Infiltration, Soil Physics, and Soil Water Storage

Infiltration is the movement of water into the soil, and the hydrology and water quality of a watershed is controlled to a large degree by the infiltration characteristics of the surface soils. Although infiltration rates in wetlands themselves are typically low, infiltration rates across the landscapes surrounding wetlands can have a strong effect on the routing of water to the wetlands. In forested humid landscapes, most precipitation that is not intercepted by the canopy infiltrates into the soil and enters the vegetative root system to be used in evapotranspiration, travels by subsurface pathways to surface waters (streams, wetlands, or lakes) found at the base of slopes, or percolates to groundwater. Infiltration rates in forested soils tend to be

TABLE 2.6
Interception as Absolute Annual Amounts (mm) and as Percentage of Annual Precipitation from Various Locations Around the World

Location	Forest Type	Precipitation (mm)	Interception (%P)	Interception (mm)
Amazon	Rainforest	2,810	9	250
Southwest India	Cashew	3,000	31	930
E. Puerto Rico, United States	Rainforest	5,750	42	2,420
S. Appalachians, United States	Mature White pine	2,030	9	180
S. Appalachians, United States	Hardwoods	2,030	12	240
S. Appalachians, United States	35-yr White pine	2,030	19	390
S. Appalachians, United States	10-yr White pine	2,030	15	300
S. I. New Zealand	Mixed forest	2,600	24	620
N. Appalachians, United States	Hardwoods	1,300	13	170
SE United Kingdom	Corsican pine	790	35	280
Norfolk, United Kingdom	Mixed pines	600	36	220
Wales, United Kingdom	Sitka spruce	1,870	27	500
Holland	Oak forest	310	22	70
S. Scotland, United Kingdom	Sitka spruce	1,600	30	480
Northumberland, United Kingdom	Mature Sitka spruce	1,000	49	490
Northumberland, United Kingdom	Sitka spruce—pole timber	1,000	29	290
S. Scotland, United Kingdom	Sitka spruce	970	32	310
Scotland, United Kingdom	Sitka spruce	2,130	28	600
NE Scotland, United Kingdom	Scots pine	640	42	270

SOURCE: Adapted from Dingman (1994); data from multiple published sources.

very high, higher than all but extreme rainfall or snowmelt rates, so the hydrology of forested hillslopes tends to be dominated by subsurface processes. Human land-use activities that compact or denude soils reduce infiltration rates, often reducing them so much they are exceeded by commonly experienced rainfall or snowmelt rates. When rainfall or snowmelt rates exceed infiltration rates, the excess water runs off the soil surface, rapidly carrying sediment and contaminants to surface waters and increasing storm flows. Maintenance of good hydrologic and water-quality conditions in surface waters is largely a matter of maintaining high infiltration rates.

The physics of infiltration are very complicated. Infiltration rates in soils are affected by soil physical properties (porosity, structure, and texture, discussed above), antecedent moisture content, the amount of `vegetative detritus on the soil surface, vegetation, layering of soils, vertebrate and invertebrate activity in the topsoil, landscape position, groundwater dynamics, and even air temperature. For given soil conditions, the potential infiltration rate decreases asymptotically over time during a wetting event (Fig. 2.8).

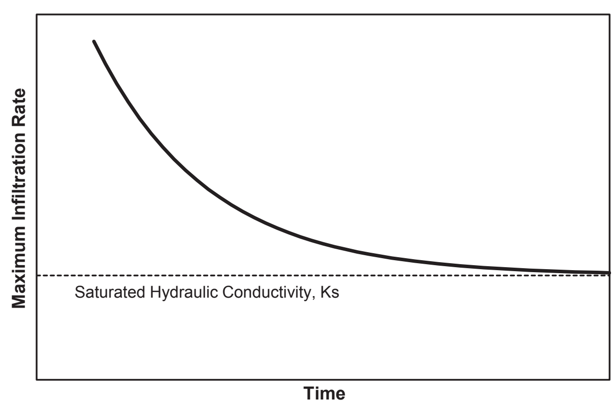

FIGURE 2.8. General behavior of maximum infiltration rates over time. When precipitation rates exceed the maximum infiltration rate an any time, overland flow occurs.

Infiltration is driven by two forces: gravity and soil capillarity. Gravity remains constant during infiltration, but soil capillary forces pulling water into the soil diminish as the soil gets wetter. The shape of the maximum infiltration rate curve matches the shape of the curve of total force (gravity + capillarity) driving infiltration over time (Fig. 2.8).

The presence or absence of vegetative detritus at the soil surface strongly affects infiltration rates. Vegetative detritus protects soils from the kinetic energy of raindrops. When vegetative detritus is removed, raindrop energy breaks up surface peds, and the resulting smaller soil particles clog the surface pores, creating a crust on the soil surface. The final infiltration rate of a crusted soil is usually a small fraction of the final infiltration rate for the same soil with vegetative detritus in place.

Soils absorb water in the same way that sponges and paper towels absorb water. A water molecule is composed of two hydrogen atoms attached by covalent bonds to a single oxygen atom. The hydrogen atoms are not arranged symmetrically on the water molecule; rather, the water molecule features a 104.5° bond angle. Therefore the water molecule is *polar,* with the oxygen atom displaying negative charge and the hydrogen atoms displaying positive charge. The polarity of the water molecule accounts for the high surface tension of water, as water molecules tend to form weak chains of molecules attracted by *cohesion* of the negative side of one molecule to the positive side of another molecule. *Capillarity* is driven by the adhesion of polar water molecules to charged surfaces and the cohesion of polar water molecules to one another. Capillarity is really electromagnetic attraction. Water adheres to soils because soil particles tend to feature negatively charged surfaces. If soil particles were uncharged, the root zone would not be able to hold water, and plant life would exist only on the margins of surface waters.

Mineral soil wetlands usually feature slowly moving surface waters that allow the settling of clays, silts, and fine organic particles, so very fine-textured soil layers tend to develop on the bottoms of wetlands. Such fine-textured soils create low infiltration rates and impede the loss of wetland surface water to the underlying groundwater system. Organic soil wetlands or peatlands develop as a result of vegetation productivity being greater than decomposition over time. Generally lower soil layers have more time to decompose and are of finer texture, leading to slower drainage. In essence, once a wetland forms, it evolves in ways that increase its wetness or hydroperiod.

Understanding the dynamics between soil water movement and soil characteristics requires a basic knowledge of soil's physical properties. Some of the major soil characteristics that hydrologists and wetland scientists consider with respect to soil hydrologic behavior are porosity, bulk density, texture, and the soil moisture release curve or moisture characteristic curve. The soil matrix is composed of mineral and organic particles, air, and water, and the relative amounts of air and water are constantly shifting as soil wets and dries. Soil texture strongly affects water storage and transmission properties of a soil (Fig 2.4). Water in soil moves from areas of high energy to areas of low energy. Many forms of energy, such as kinetic, thermal, and osmotic, are negligible in most soil and groundwater flow situations. Typically, only potential energy and pressure energy are considered in soil water and groundwater movement:

$$\text{Soil water energy} = \text{potential energy} + \text{pressure energy} \tag{2.4}$$

The pressure in a standing column of water increases linearly with depth, but the total energy stays constant because the potential energy decreases as the pressure energy increases. Hydrologists and hydraulic engineers use the concept of *hydraulic head,* which is the energy per specific weight of fluid and has units of length, to describe the amount of energy in water (Equation 2.5). For example, the pressure head below a 10 m water column is 10 m:

$$\text{head} = \frac{\textit{energy}}{(\text{specific weight})} \tag{2.5}$$

$$\text{total hydraulic head, } H = \text{potential head} + \text{pressure head} \tag{2.6}$$

$$H = z + \Psi$$

Below the water table, water molecules are under compression, and pressure head is considered to be positive. In unsaturated soil, water molecules are under tension as they are being pulled downward by gravity and pulled toward soil particles by electrostatic attraction. Pressure head is negative for unsaturated soils. The drier the soil, the greater the negative pressure in the soil. The relationship between a soil's moisture content and its pressure is called the *soil moisture characteristic curve* or the *soil moisture release curve.*

Soil texture, structure, and organic matter content affect the shape of the soil moisture release curve. For example, sandy soils tend to drain quickly because of their higher proportion of large pores, while clayey soils tend to drain slowly because of their higher proportion of small pores (Figs. 2.9 and 2.4). Moderate amounts of humus promote the development and maintenance of soil structure and the associated macropores (e.g., Fig. 2.4), while high amounts of organic matter can seal the soil and impede water movement, as discussed previously. Neither sandy nor clayey soils are ideal for plant growth because of their water-holding characteristics. Sandy soils drain too quickly for plant roots to access much water, and clayey soils hold water so tightly that plants cannot access much of the water held in the small pores of clay soils. Soil scientists use the concept of *field capacity* to describe the water held in a soil after rapid gravitational drainage, and field capacity corresponds to water held at approximately –1.0 m negative pressure head (or, as soil scientists would say, 0.1 bars of tension). Sandy soils hold relatively little water at field capacity whereas clays are nearly saturated at field capacity. At the other end of the soil moisture release curve, soil scientists

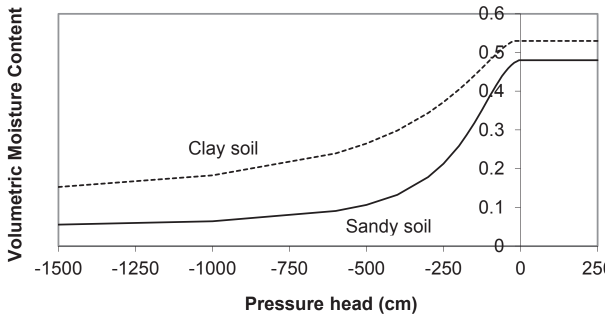

FIGURE 2.9. Representative soil moisture release curves.

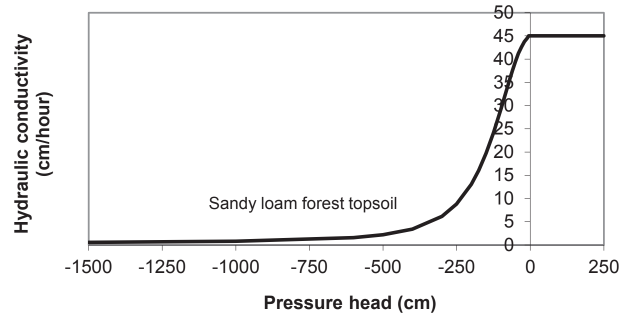

FIGURE 2.10. Typical hydraulic conductivity versus pressure head relationship.

define *wilting point* as the water held at approximately –150 m of pressure head (soil scientists would call this 15 bars of tension), which is about the maximum tension plants can apply to extract water from soil. A sandy soil will be nearly dry at the wilting point, whereas a clay soil will still hold significant amounts of water. The difference in water storage between field capacity and wilting point is often called *dynamic soil storage* or *maximum plant-available water*. Loamy soils, composed of moderate amounts of sands, silts, and clays (Fig. 2.4), have the greatest dynamic soil storage and feature optimal moisture dynamics for plant growth.

Soil water moves from areas of high hydraulic head to areas of low hydraulic head by the route of least hydraulic resistance, according to Darcy's law.

Darcy's law in one dimension for saturated soils:

$$Qx = -KxA\frac{\delta H}{\delta x} \tag{2.7}$$

Q_x = volume of flow per unit time (L^3/t),

K = Saturated hydraulic conductivity of the soil (L/t),

A = cross sectional area of flow (L^2), and

$\frac{\delta H}{\delta x}$ = hydraulic gradient, the rate of change of head over distance (unitless ratio).

Darcy's law says that the volume of water per unit time that moves from one region to another in soils is proportional to the area of flow and to the *hydraulic gradient,* defined as the rate of change of hydraulic head over distance. The hydraulic gradient can be considered the driving force for soil and groundwater movement. The flow of water in soil is also proportional to the *hydraulic conductivity* of the soil, which defines the rate of water movement across a flow cross section under a unit hydraulic gradient. The hydraulic conductivity is an intrinsic flow property of soils and bedrock that is affected by porosity, texture, structure, and macropore networks. Hydraulic conductivities of natural soils and rock vary over 12 orders of magnitude (Freeze and Cherry 1979). Spatial variations of hydraulic conductivities strongly control the hydrology of hillslopes and wetlands.

When a soil is saturated and the pressure head is positive, hydraulic conductivity is constant and referred to as the *saturated hydraulic conductivity*. Hydraulic conductivity rapidly decreases as soils dry. For unsaturated conditions, there is a relationship between hydraulic conductivity and pressure head, as shown in Figure 2.10. As do the moisture release curves, conductivity vs. pressure curves vary with soil porosity, texture, structure, and macropore networks. Using these relationships between moisture content, hydraulic conductivity, and pressure, Darcy's law can be generalized for unsaturated flow (Sidebar 2.1).

In many hydrologic investigations, soil and groundwater movement can be simplified to one- or two-dimensional systems, but in reality—including in wetlands—groundwater moves in three dimensions. Furthermore, some wetlands may receive or discharge soil water as both unsaturated and saturated flow, making simple hydrologic characterization difficult. Three-dimensional saturated/unsaturated flow problems are analyzed by a differential equation called *the Richards equation* (Richards 1931). The Richards equation is a generalization of Darcy's law coupled with the continuity equation as follows:

$$\frac{\delta}{\delta Xi}(Ki(\Psi)\frac{\delta H}{\delta x}) = C(\Psi)\frac{\delta \Psi}{\delta T} \tag{2.8}$$

$$C = \frac{\delta \theta}{\delta \Psi}$$

θ = volumetric moisture content.

The Richards equation simply says that, at any point and time, flow will move in the direction where the product of the conductivity and the hydraulic gradient are greatest. For most transient analyses of groundwater flow, the Richards equation cannot be solved analytically, so finite-element or finite-difference models of the Richards equation are used to describe soil and groundwater flow systems. Finite element and finite difference models are different techniques used to transform differential equations into sets of algebraic equations that can be solved iteratively on computers (see examples in Winter 1999).

When considered in the light of soil physical proper-

SIDEBAR 2.1 DARCY'S LAW GENERALIZED FOR THREE DIMENSIONS AND UNSATURATED AND SATURATED FLOW

Darcy's law can be modified to describe both saturated and unsaturated flow by substituting conductivity as a function of pressure head into the equation.

$$Qx = -K_x(\psi)\, A\, \frac{\delta H}{\delta x}$$

The terms are the same as in Equation 2.7, except that now the hydraulic conductivity is not a constant but is a function of the pressure head in the soil. Dividing both sides of Equation 2.7 by the cross-sectional area of flow (A) produces the equation for *specific flux,* or *Darcy velocity, q*. The specific flux is the average flow rate across the entire flow cross section. However, water cannot flow through the soil particles themselves, only through the wetted portion of the pore space. The *pore velocity* is the average velocity of water molecules and can be calculated by dividing the specific flux by the volumetric moisture content.

$$qx = -K_x(\psi)\, \frac{\delta H}{\delta x}$$

qx = specific flux or Darcy velocity in the x direction, the average rate of water flow assuming water flows through the entire flow cross section (which it does not). Specific flux has units of velocity (L/t).

pore velocity = $q/(\theta_v)$

θ_v = volumetric moisture content = V_w/V_s

The pore velocity describes the average rate of movement of a solute, such as a contaminant or a tracer, being carried by soil water.

ties and the Richards equation, the temporal dynamics of maximum infiltration rates and the variability of infiltration rates across soil and cover types are more easily understood. At the beginning of a precipitation or ponding event, there is a large pressure head gradient between the just-wetted soils at the surface and the drier soils below. In essence, capillary forces pull water from the wet surface into the drier soils below. The sharp interface between high moisture contents and low moisture contents is called the *wetting front*. As the wetting front advances deeper into the soil, the pressure gradient at the surface approaches zero. At this point, flow at the soil surface is driven only by gravity. In other words, infiltration is driven by the potential head gradient in the vertical direction, which is equal to one when there is no vertical variation in moisture content or pressure. If the rainfall or snowmelt rate is less than the saturated hydraulic conductivity of the soil, then equilibrium will be reached when the soil reaches a moisture content and a pressure at which the hydraulic conductivity equals the input rate. If the rainfall or snowmelt rate exceeds the saturated hydraulic conductivity of the soil, the infiltration rate will approach the saturated hydraulic conductivity and the excess water will flow over the soil surface (see the following section on Horton overland flow). Maintaining high infiltration rates and minimizing surface runoff involves keeping an organic cover on the soil; avoiding soil compaction; and maintaining a healthy soil ecosystem with plants, macroinvertebrates, and even small mammals that create macropore networks with high flow capacities.

Overland Flow

When precipitation rates exceed soil infiltration rates, the excess runs over the ground surface, creating *surface runoff*, called *Horton overland flow* by hydrologists (Horton 1933). Over short distances, this surface runoff will travel as a dispersed film, but after traveling tens of meters, it will concentrate into small channels as a result of microtopographic effects and because of scour caused by the flow itself. Horton overland flow is rare in humid forested environments because vegetative detritus, high soil organic content, and biological activity in the topsoils (due to roots, invertebrates, small mammals) all combine to create very high infiltration rates. Horton overland flow may become common when soils are compacted and vegetative detritus is removed from the surface. Thus, it is common on row-crop agricultural fields, heavily grazed pastures, residential lawns, and obviously pavements and rooftops. Horton overland flow is important to wetlands in agricultural, suburban, or urban environments. Horton overland flow scours and transports sediment and any other pollutants found on the ground surface, creating water-quality problems in the receiving waters. At the extreme, Horton overland flow carves rills and gullies and then deposits large amounts of sediment on floodplains and in streams and wetlands (Richter and Markewitz 2001).

The fundamental hydrologic shift engendered by watershed urbanization is the frequent and widespread occurrence of Horton overland flow on pavements, rooftops, and compacted soils that have little to no infiltration capacity. This alteration of the infiltration capacity of the landscape results in higher peak flow rates and stormflow volumes, and also produces large stormflows in the summer and early fall when formerly forested basins normally produce no significant runoff events because of dry soil conditions. Overland flows from urban areas typically carry high concentrations of fertilizers, pesticides, oils and grease, sediment, and heavy metals; overland flows from row-crop agricultural areas typically carry fertilizers, pesticides, and sediment. Thus overland flow is an important source of contaminants to wetlands.

Evapotranspiration

Hydrologists usually lump *evaporation*, water moving from a water body to the atmosphere, with *transpiration*, water moving from plant tissue to the atmosphere, into a single metric called *evapotranspiration*. Accurate monitoring and estimation of evapotranspiration is difficult and one of the most problematic issues in water budgeting. In most analyses of wetland water budgets, however, it makes sense to consider evaporation and transpiration separately.

Evaporation is the change of state of liquid water, either from an open water body or from the ground, into vapor and the transfer of this vapor to the atmosphere. Evaporation is an energy-driven process that occurs when molecules of liquid water attain enough kinetic energy to overcome surface tension and escape from the water surface.

Most of the energy that drives evaporation comes from solar radiation and sensible heat transfer (exchange of heat) from the atmosphere. As a result, evaporation varies by season and time of day. The energy it takes to evaporate a gram of water is quite large and is known as the *latent heat of vaporization*, which is equal to about 600 cal/g. On a small scale, evaporation and condensation are always occurring simultaneously in the air over a water layer. The evaporation rate depends on the humidity in the air and also on the wind speed. As the humidity of the air increases, so does the partial pressure of water vapor in the air, and it becomes more difficult for water to evaporate. At a given temperature and solar radiation, evaporation is greatest on a windy, dry day and lowest on a still, humid day. In other words, the evaporation increases with increasing vapor pressure deficit:

Vapor Pressure Deficit: (2.9)

$$VPD = e_s - e_a = e_s (1 - RH)$$

where

e_s = the maximum, or saturated vapor pressure,

e_a = the actual vapor pressure,

RH, relative humidity = $\frac{e_a}{e_s}$

Basically, this equation means that drier air has a greater moisture deficit. In 1802, English chemist John Dalton developed mass transfer equations of the following form for predicting free water evaporation (Dingman 1994):

$$E = b\, v_a (e_s - e_a) \tag{2.10}$$

where

E = evaporation in cm/day,

v_a = wind speed 2 m above water surface, cm/s, and

b = empirical constant (depends on ht. of wind and vapor pressure measurements).

Evaporation from an open water body is called *potential evapotranspiration* (PET) because it is not limited by water availability. PET is typically measured using a United States Weather Bureau Class A pan (Shaw 1988), which is a circular aluminum pan 1.22 m diameter and 25.4 cm high with an operating water depth between 17.5 and 20 cm. Annual PET varies relatively predictably across the United States (Farnsworth and Thompson 1982), and varies relatively little from year to year (Dingman 1994). PET data for particular areas can be obtained from the National Weather Service and from many state climatological data programs.

Evaporation from soils is complicated because water supply is limited and water is bound to soil by capillary forces. After the first few millimeters of water evaporate from soil near the surface, vapor must diffuse out of the pores. By setting up pressure gradients between dry surface soils and wetter soils below, evaporation can cause upward vertical unsaturated flow. Evaporation from soil is a very slow process after the top few centimeters become dry. If the soil is covered by a layer of vegetative detritus, then evaporation from soil is further inhibited. Therefore, water loss from the subsurface soils to the atmosphere occurs primarily through plant *transpiration*.

Plants use atmospheric carbon dioxide (a trace component of the air with a concentration of about 400 ppm and rising) as the source of carbon for building plant tissue. To access carbon dioxide, leaves of plants contain small gas chambers that exchange gas with the atmosphere through openings called *stomata*. When a stoma is closed, the air in the gas compartment becomes nearly saturated with water

TABLE 2.7
Gross Watershed Water Budgets for Various Locations in the United States

Location	Precipitation (cm)	Streamflow (cm)	AET (cm)
Atlanta, Georgia	130	40	90
Seattle, Washington	100	50	50
Olympic Mountains, Washintgon	300	250	50
Tucson, Arizona[a]	30	0	90

[a] Actual evapotranspiration (AET) in and around Tucson, Arizona, exceeds precipitation because of irrigation using groundwater or water imported from the Colorado River.

vapor; when the stoma is open, this high humidity air is exchanged with lower humidity air, causing a net loss of water. Stomata close when cell wall turgor pressure drops because of water deficits in leaves. This transfer of water from leaves to the atmosphere is called *transpiration,* and it is a necessary byproduct of the plant's need for carbon dioxide. Transpiration also drives the movement of water from the roots to the canopy and thus allows delivery of essential soil nutrients to plant tissue. Like evaporation, transpiration rates are partly controlled by vapor pressure deficit and wind, but transpiration is also controlled by the supply of water in the soil, leaf area index, plant species, solar radiation, and air temperature. Because of their ample water supplies and typically lush vegetation, transpiration can be a major source of water movement from wetlands.

The actual evapotranspiration (AET) from the landscape on any day is almost always less than the PET for that day. On an annual basis, AET is typically in the range of 50% to 90% of PET, with higher ratios of AET:PET in wet climates and lower ratios in arid climates. One simple way to estimate AET for an open water surface is by multiplying PET measured in pans by the pan coefficient, which is the ratio of annual AET to annual PET as follows:

$$AET = Cp \times PET \qquad (2.11)$$

where

Cp = pan coefficient = annual AET/annual PET

Annual AET can be estimated using a simplified basin water budget. Assuming that trans-watershed human and groundwater exchanges are negligible when compared to surface flows leaving the basin, then long-term average annual precipitation and runoff data can be used to estimate annual AET as follows:

$$P = AET + R \qquad (2.12a)$$

or

$$AET = P - R \qquad (2.12b)$$

where

AET = average annual actual evapotranspiration in cm,

P = average annual precipitation in cm, and

R = average annual runoff expressed as a depth in cm.

This equation says that, over the long term (a long-term record renders negligible any changes in soil or surface water storage), precipitation becomes either evapotranspiration or runoff. The partitioning of precipitation into evapotranspiration or runoff varies greatly with climate type (Table 2.7).

Predicting and modeling evapotranspiration based on physical principles is difficult because of the complexities of evapotranspiration processes, differences in water usage between vegetative communities, the strong control exerted by soil moisture availability, and the temporal dynamics of the physical forcing functions (humidity, temperature, solar insolation, wind, etc.). The more physically realistic evapotranspiration equations are based on energy budget considerations. An energy budget model is based on measuring all major incoming and outgoing energy fluxes as well as changes in energy storage, and then solving for the energy used in evapotranspiration. The Penman-Monteith equation (Monteith 1965) combines energy budget and mass-transfer concepts and also accounts for canopy conductance, and is commonly used for estimating actual evapotranspiration. The Penman-Monteith equation was modified from the Penman equation (Penman 1948) for predicting evaporation from lakes. These equations ignore smaller energy flux terms such as changes in heat storage and energy carried by water flowing into and out of the water body. With calibration of the conductance terms, the Penman-Monteith equation has accurately estimated evapotranspiration

SIDEBAR 2.2 THE THORNTHWAITE EQUATION, AN INDEX MODEL FOR ESTIMATING POTENTIAL EVAPORATION BASED ON AIR TEMPERATURE

The Thornthwaite equation predicts monthly PET as follows:

$$Et = 1.6\left[\frac{10Ti}{I}\right]^{a}$$

Et = potential evapotranspiration in cm/month

Ti = Mean monthly air temperature (degrees C)

$$I = \text{annual heat index} = \sum_{i=1}^{12}\left[\frac{Ti}{5}\right]^{1.5}$$

$$a = 0.49 + 0.179I - 7.71\times10^{-5}\, I^2 + 6.75\times10^{-7}\, I^3$$

The only data needed for the Thornthwaite equation are average monthly air temperatures, making this a useful equation when more complete climate data are not available.

in many environments (e.g., Calder 1977, 1978; Lindroth 1985; Dolman et al. 1988; Stewart and Gay 1989; Lemeur and Zhang 1990). The Priestley-Taylor equation is similar to the Penman-Monteith equation, but also includes heat flux from the ground and a crop and climate calibration factor. It requires data on air temperature, vapor pressure, net radiation, and soil temperature. The Priestley-Taylor equation has been found to accurately estimate evapotranspiration in a variety of wetland, forest, and arid vegetated environments (e.g., Drexler et al. 2004; Lu et al. 2005; Chulow et al. 2012).

The Penman-Monteith, Priestley-Taylor, and other energy-budget models require a lot of data that may not be available for a given site and time period. The empirical Thornthwaite model (Sidebar 2.2) of potential evapotranspiration is a commonly used index model that uses only temperature data to predict PET (Thornthwaite and Hare 1965). The idea of an index model is that hot days typically have higher solar radiation and vapor pressure deficit, and thus higher PET, while cool days typically have lower solar radiation and vapor pressure deficit, and thus lower PET.

Another relatively simple way to estimate AET is with a soil-factor (F_s) and crop-factor (F_c) model that accounts for soil moisture and plant conditions.

$$AET = F_s\, F_c\, PET \qquad (2.13)$$

The soil factor varies between 0, when the soil is at wilting point, and 1, when the soil is at field capacity. Various equations have been developed to relate the soil factor to soil moisture content (Shuttleworth 1992), but the simplest of these equations assumes a linear relationship as follows:

$$Fs = \frac{(\theta - \theta_{wp})}{(\theta - \theta_{fc})} \qquad (2.14)$$

where

θ = current moisture content,

θ_{fc} = moisture content at field capacity, and

θ_{wp} = moisture content at wilting point.

A soil moisture accounting model can be developed to track soil moisture and AET over time. This is the way some golf courses and more sophisticated farmers optimize irrigation. The crop factor is a function of growth stage, and various empirical crop factor relationships can be found in the agricultural literature (Shuttleworth 1992).

This large but far from all-inclusive discussion of models is meant to introduce the array of approaches and simplifying assumptions used for estimating evapotranspiration (Abtew and Melesse 2012). All of these models can be used to estimate AET and PET from wetlands (Drexler et al. 2004), and the appropriate model choice depends on data availability, the sophistication of analysis, and the wetland vegetation. Evapotranspiration from a wetland differs from evaporation from open water (Mohammed et al. 2012), and the type and condition of vegetation affects evapotranspiration (e.g., Mohammed et al. 2012; Brown et al. 2010). If required data are available, the Priestley-Taylor equation has been found to estimate evapotranspiration robustly for wetlands and other vegetated environments (e.g., Lu et al. 2005; Chulow et al. 2012), but in some cases little difference has been found among the estimates of different evapotranspiration estimation approaches (e.g., Mao et al. 2002; Rosenberry et al. 2004; Masoner and Stannard 2010).

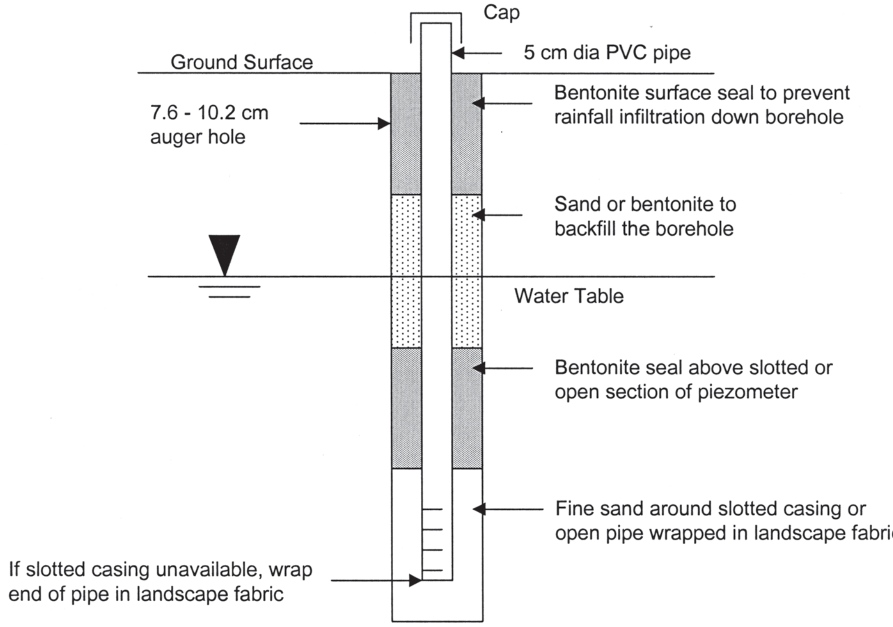

FIGURE 2.11. Well schematic.

Interflow

Interflow is shallow, lateral subsurface flow moving nearly parallel to the soil surface over an impeding soil or bedrock horizon. Interflow during or just following rainfall or snowmelt is analogous to Horton overland flow because interflow occurs when a wetting front delivers water through a soil horizon faster than the lower horizon can accept water. Hydraulic conductivities of soils tend to decrease with depth, and conductivities and other hydraulic properties may feature abrupt changes at the interfaces of soil layers. As infiltrating water reaches a soil layer that impedes percolation, such as a dense clay B-horizon, some of the soil water will begin to flow laterally downslope over the impeding horizon. During storms, interflow can occur as saturated or unsaturated flow. Interflow during storms will not occur until after the wetting front has crossed the topsoil and encountered the impeding horizon (Zaslavsky and Sinai 1981a, 1981b; Wallach and Zaslavsky 1991). Therefore, a threshold precipitation amount, typically in the range of 50 mm, is necessary before interflow becomes an important part of stormflow processes (Newman et al. 1988; Freer et al. 2002; Tromp van Meerveld and McDonnell 2006).

Interflow also occurs as unsaturated flow during drying conditions between rainstorms and thus serves to redistribute moisture and solutes from upslope to downslope (Hewlett and Hibbert 1963; Jackson 1992). The unsaturated drainage of shallow soils by interflow supports baseflows in streams and wetlands. Because interflow occurs in the upper soil layers where organic content and biological activity are high, many biochemical transformations of materials dissolved or suspended within interflow may occur. Travel times associated with interflow are on the order of hours to months, depending on slope position and soil moisture.

Percolation

Soil water that has infiltrated, avoided root uptake and transpiration, and passed through the topsoil now flows downward through the *vadose zone* driven mostly by gravity but also partly by horizontal and vertical pressure gradients developed during hillslope drying and moisture redistribution. The vadose zone is comprised of the soil and rock between the ground surface and the water table. It is usually unsaturated but may become saturated for short periods of time during precipitation events. The downward movement of soil water to the underlying water table is called *percolation*, and percolating water that reaches the water table is called *recharge*. Below ridgetops the depth to the water table may be large, so travel times for percolating water may be months, even years. Near the base of hillslopes, where the water table is near the ground surface, percolation travel times are relatively short. The vadose zone beneath the root zone, and the percolating water within it, act to buffer streamflows and wetland hydropatterns from changes in climatic conditions. During dry periods, the vadose zone stores precipitation and minimizes stormflow response; during wet periods, the percolating water supplies recharge to aquifers and helps support baseflow in streams.

Groundwater Flow

Groundwater hydrology encompasses the subsurface movement of water as both unsaturated flow in the vadose zone and saturated flow beneath the water table. Ironically, in most humid natural environments, surface water hydrology—encompassing the behavior of wetlands, streams, rivers, and lakes—is strongly associated with groundwater flow processes because surface waters receive most of their flow from various groundwater flow paths. This discussion

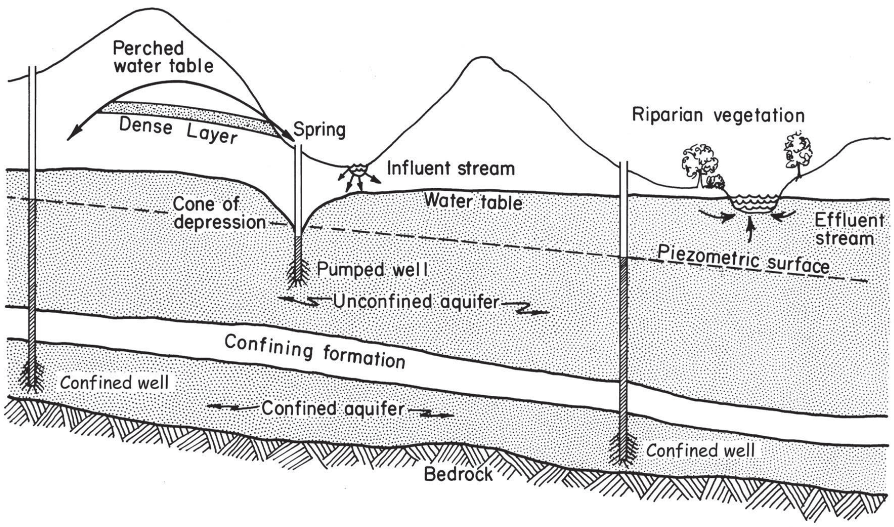

FIGURE 2.12. Aquifer types and surface water interactions. From Hewlett (1982).

of groundwater flow, however, will focus on saturated flow systems and their relationships to surface waters, particularly wetlands. Laypeople often say "water flows downhill" and residents of arid environments often say "water flows toward money," but it is most accurate to say that "water flows from high head to low head," as described by Darcy's law. Surface water bodies, such as wetlands, streams, and rivers, are located at local low points in the landscape, and in humid environments they generally act to drain groundwater from the surrounding landscape.

Hydraulic head of groundwater is measured using wells or piezometers (Fig. 2.11). The elevation of the equilibrium water surface in a well is the total hydraulic head associated with the screened section of the well, and it defines the water table at that point. The *water table* is the three-dimensional surface defined by water standing in wells in that area. Above the water table, soil water is held in tension, and thus the pressure head is negative. Below the water table, ground water is under positive pressure. Another definition of the water table is the surface at which water pressure (or pressure head) equals zero. Water tables are partly controlled by surface topography and tend to form a muted reflection of the surface topography.

The Earth is approximately 4.6 billion years old, and the hydrologic cycle has been running for about 4 billion years. Consequently, the fractured surface mantle is full of circulating water down to a depth of about 10 km. In humid environments, this groundwater is constantly recharged and leaks into the surface water system from hillsides. In arid environments, saturated groundwater systems occur at depths where evaporation losses and drainage to local surface water bodies are minimal. The movement of groundwater and its exchanges with surface waters occur at all spatial and temporal scales. Wetlands generally occur where the saturated groundwater system approaches the soil surface because of topographic, geologic, or soil conditions.

An *aquifer* is any geologic medium that will transmit usable quantities of water to a well, so aquifers have relatively high saturated hydraulic conductivities and porosities. Conversely, an *aquitard* is a geologic medium with limited transmission of water, and an *aquiclude* is a geologic medium that essentially moves no water because either the porosity or the saturated hydraulic conductivity are practically zero. The low-conductivity Histosols that sometimes form below wetlands typically act as thin aquitards.

There are two general types of aquifers: surficial (unconfined) and confined, although many aquifers are semi-confined (Fig. 2.12). In a *surficial aquifer*, the water table is usually open to the atmosphere through the air-filled pore spaces in the vadose zone. The entire overlying landscape is a recharge zone for surficial aquifers. When a surficial aquifer is recharged by percolation, the zone of saturation expands to the same extent that the water table rises. Surficial unconfined aquifers continually drain to local surface water bodies and provide streamflow between storms. As water drains out of the surficial aquifers, water tables drop, and the rate of discharge from aquifers to streams diminishes, thus lowering baseflow between storms.

A *confined aquifer* is sandwiched between two aquitards or aquicludes. Water fully saturates a confined aquifer, and the zone of saturation does not change as pressure changes in the aquifer. Water levels of wells drilled into a confined aquifer form a *potentiometric surface*. Where the potentiometric surface rises above the ground surface, the aquifer is artesian, and wells will flow without the aid of a pump in these areas. Typically, a confined aquifer outcrops somewhere on the landscape, meaning that the geologic strata bearing the aquifer contacts the land surface at some location. These outcrop areas are *recharge areas* for the confined aquifer because recharge from above the aquifer is limited by the overlying aquitard. Usually, hydrologic connections between confined aquifers and surface water are limited and damped, but in some cases, rivers cut into confined aquifers and thus interactions are direct and rapid.

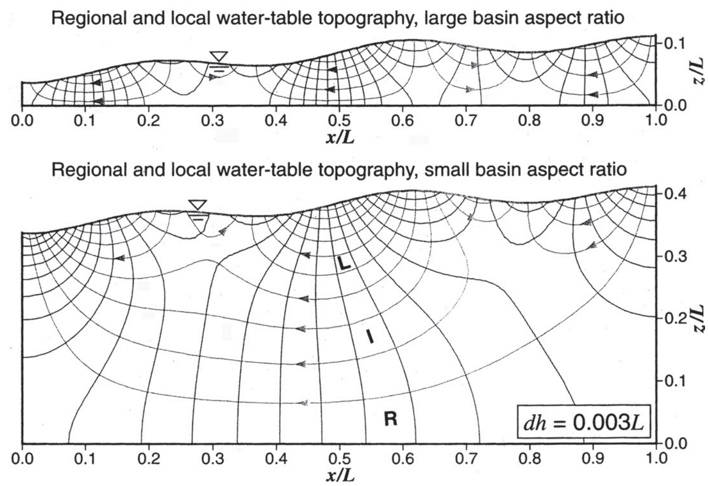

FIGURE 2.13. Local and regional groundwater flow nets. From Hornberger et al. (1998).

At baseflow, streams and rivers typically flow at velocities ranging from 0.3 to 1.5 m/s. Even a river 300 km long would only take from 2 to 11 days to completely drain if there were no continuous inputs of water. Groundwater is what sustains streams, rivers, and many wetlands between precipitation or snowmelt events.

Groundwater flow paths are complicated and three-dimensional (Fig. 2.13). Recharge water that reaches the water table below ridgetops takes a deep and circuitous path to streams that may take many years. Ridgetop recharge water may not flow to the local surface water network but rather may enter the regional flow network and travel to the ocean or a lake without ever entering the stream and wetland system. Recharge water that reaches the water table near a stream, on the other hand, will travel a relatively fast (on the order of days or weeks) and shallow route to the stream. Consequently, geochemical transformations of groundwater vary by flow path. For example, the chemistry of shallow groundwater near the valley floor is affected by rapid microbial transformations and may undergo denitrification processes. Deep groundwater travels in a medium with little organic carbon, and biological interactions with groundwater chemistry are minimal.

Variable Source Area Runoff or Saturated Surface Runoff

Where water tables lie close to the ground surface, such as in low-lying areas near streams, in low portions of floodplains, and around wetlands, water tables may rise to the ground surface during precipitation. When the soil becomes saturated, additional precipitation runs across the soil surface to surface waters, and runoff from this mechanism is called *variable source area runoff*, or *saturated surface runoff* (e.g., Betson 1964; Hewlett and Hibbert 1967; Ragan 1967; Dunne and Black 1970a, 1970b; Dunne et al. 1975). These runoff source areas are "variable" in that the saturated areas expand and contract with climatic and groundwater conditions. Variable source areas are fed by groundwater and interflow, and the water table position in these areas is partly controlled by water levels in the nearby surface waters. Variable source areas are responsible for most of the stormflow from undeveloped forested watersheds in humid climates. From a hydrologist's viewpoint, most wetlands are variable source areas.

Summary of Hillslope Flow Process

These categorizations and descriptions of flow processes are conceptually useful, and they allow for ready and simple explanations of how hydrology differs among mountain hillsides, Piedmont hillsides, Coastal Plain hillsides, and glacial landscapes. With basic information on soils and topography, a hillslope's dominant hydrologic processes can be inferred (Fig. 2.14). Furthermore, this simple reductionist view of hillslope flow processes can guide the development of a wetland water budget and illuminate controls on a wetland's biogeochemistry.

As soon as precipitation hits the soil surface, it is connected by soil water and groundwater to wetlands, streams, lakes, and rivers. There exists a continuous dynamic hydrologic connection between the soil surface and surface waters large and small, including "Waters of the United States." In *Rapanos vs. The United States* (see Chapter 8), the Supreme Court decided that federal jurisdiction under the Clean Water Act extended only to surface water with significant nexus to "Waters of the United States." From a hydrologic perspective, any such distinction is arbitrary as there is continuous connectivity of water, solutes, and energy through the terrestrial portion of the hydrologic cycle.

No unmanaged hydrologic system is static. Streamflow, groundwater levels, soil moisture, and wetland water levels are always either increasing because of recent precipitation or decreasing because of drainage and evapotranspiration (Fig. 2.15). Because of local hydraulic controls, such as beaver dams, some wetlands may experience relatively long periods of time with low water-level fluctations, but inflows and outflows are always in transition.

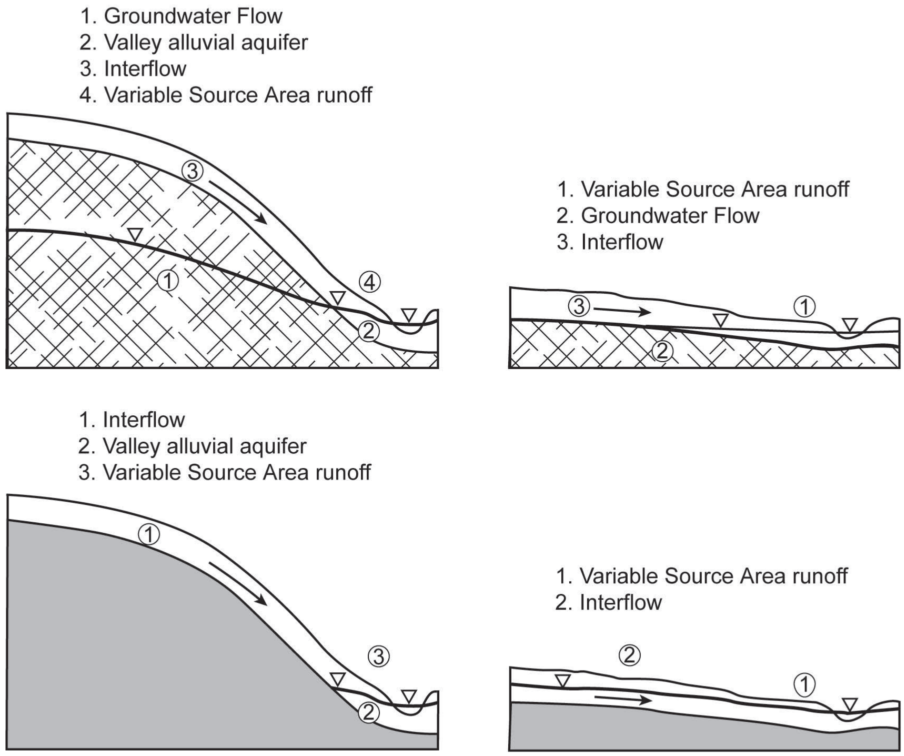

FIGURE 2.14. Examples of how hillslope topography, soils, and lithology affect the relative importance of possible hillslope flow pathways.

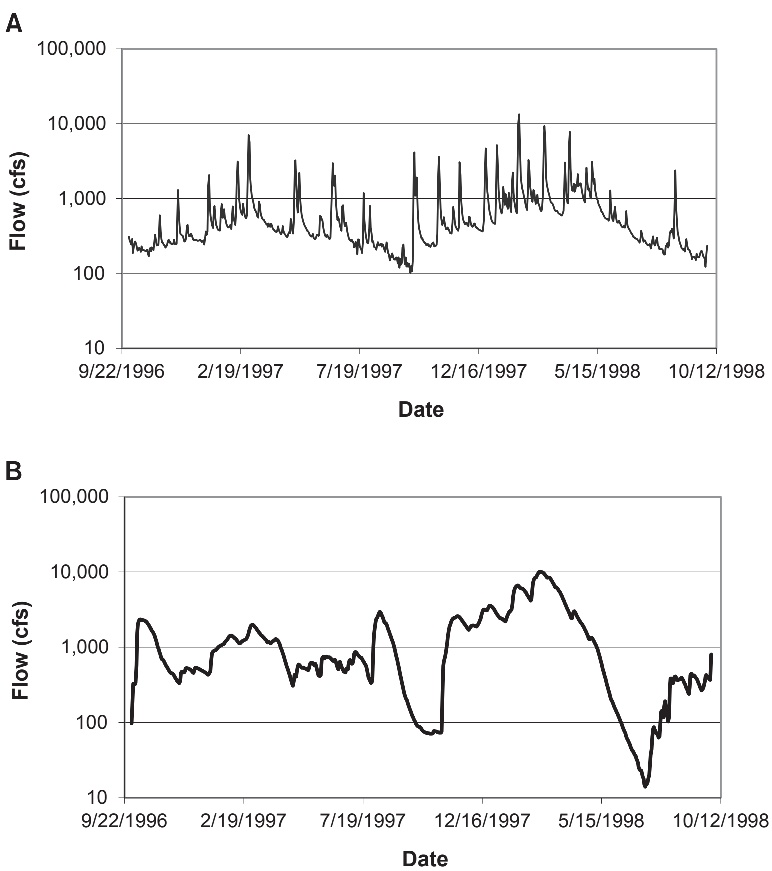

FIGURE 2.15. Hydrographs for two rivers in Georgia for water years 1997 and 1998. One is a free-flowing Piedmont river and the other is the outflow of the Okefenokee Swamp complex in the lower Coastal Plain. (A) Hydrograph of the Middle Oconee River in Athens, Georgia, which drains a basin of 392 square miles of mostly forested uplands with few floodplain wetlands. (B) Hydrograph of the Suwannee River, which drains 1,260 square miles dominated by the Okefenokee Swamp. Flows are in cubic feet per second from USGS gages 02217500 and 02314500.

Wetland Water Budgets

Quantitative or semi-quantitative water budgets provide useful insight into the hydrodynamic and geochemical behavior of a wetland. However, an accurate quantitative water budget is time-consuming, expensive, and usually beyond the scope of most wetland studies. Nevertheless, relatively simple hydrologic data collection and analysis can provide a rough and useful quantification of a wetland's water budget. Water budgeting is based on the simple physical principle of *conservation of mass*, or *continuity*, which says that the mass (or volume) of inputs to a system must equal the mass (or volume) of outputs plus the change in internal system storage over any period of time.

The continuity equation:

inputs = outputs + changes in storage,

When applied to a freshwater wetland, is:

$$P + G_{in} + OF_{in} + SF_{in} + OBF_{in} = ET + G_{out} + SF_{out} + OBF_{out} + dV \quad (2.15)$$

where

P = volume of precipitation falling on wetland,

G_{in} = volume of groundwater flow into wetland,

G_{out} = volume of groundwater flow leaving wetland,

OF_{in} = volume of overland flow into wetland,

SF_{in} = volume of streamflow into wetland,

SF_{out} = volume of streamflow leaving wetland,

OBF_{in} = volume of overbank flow into wetland,

OBF_{out} = volume of overbank flow leaving wetland,

ET = evapotranspiration, and

dV = change in volume of water stored in wetland.

Depending on the goal of the investigation, a wetland water budget can be more finely or more coarsely partitioned. Groundwater flow can be separated into interflow and surficial aquifer flow, and overland flow can be separated into variable source area runoff and Horton overland flow from upslope. Streamflow and overbank flow could be considered separately or together. The first step in creating a water budget is to identify the possible water inputs and outputs (Table 2.8). The second step is to decide the time scales of interest, as this will directly affect the choice of monitoring and analytical methods. The third step is to identify and collate the climate and streamflow data—such as amounts; seasonality; and variability in precipitation, evapotranspiration, streamflow, and water table levels—available for characterizing regional hydrology.

TABLE 2.8
Possible Sources and Losses of Water for Wetlands

Possible Water Sources	Possible Water Losses
Precipitation	Evapotranspiration
Streamflow	Streamflow
River overflow during floods	River return flow following floods
Groundwater	Groundwater
Variable source area runoff	Tides
Horton overland flow	Human withdrawals
Tides	
Human inputs	

Water budgets can be applied conceptually with general information about climate and landscape conditions. It is obvious that the water budget of a wetland located on a perennial stream is dominated by streamflow input and output, and such a wetland will stay saturated year round. A desert playa wetland generally receives water only from precipitation, and the wetland is inundated only for a short period of time until evapotranspiration removes the water. A floodplain backwater swamp may be fed in most seasons and most years by the regional water table and by hillslope interflow, but during a flood period the wetland's water may be completely exchanged with river overflow water. Depending on the river and floodplain characteristics, it may be months or years between flood events that reach the wetland. A wetland's water budget has direct implications for biogeochemical cycling, as discussed in Chapter 4.

Outside of the tropics, there is always strong seasonality to evapotranspiration demands. Therefore, even if monthly precipitation is relatively constant, wetland water budgets and water levels usually exhibit seasonality. When seasonality of precipitation or snowmelt also affects a wetland's water budget, the tendency for seasonal behavior is enhanced. The information provided by a water budget depends on the time scale of the data and the analysis.

Hydropatterns

The *hydropattern* is the typical or average behavior of a wetland's water-level time series. Wetland hydropatterns are important ecologically because most aquatic organisms are associated with the water column and near-surface soils, and their life histories must be synchronized to the inundation periods of a wetland. In general, the hydropattern is "the seasonal pattern of the water level of a wetland and is

TABLE 2.9
Some Metrics for Characterizing Wetland Hydropatterns

Metric	Notes
Duration of inundation period	Usually expressed in months or weeks, this metric is the average annual duration of surface saturation.
Start and end dates of surface inundation	Wetlands may have equivalent durations of surface inundation, but these periods may occur at different times of year. The temporal relationship of the inundation period to the life histories of plants, macroinvertebrates, waterfowl, and amphibians has obvious repercussions for wetland ecology.
Stage duration curve	This metric is a curve that relates the fraction of time the wetland water level equals or exceeds a given stage. The statistical analog in stream hydrology is the flow-duration curve. This curve does not yield information about the number of times water level rises and drops across a specific level.
Stage excursion frequency	This metric is a histogram showing the number of events per year during which any given range of stages is equaled or exceeded.
Stage recurrence curve	This curve describes the annual probability of exceedance or the average return period of all observed peak annual stages. The statistical analog in stream hydrology is the peak flow recurrence curve.
Mean water level fluctuation	The idea behind the mean water level fluctuation is to characterize how much the water level rises and falls—that is, the "flashiness" of the water levels. The highest and lowest water levels are recorded for specified time intervals, usually two weeks or a month. The metric is not independent of the sampling frequency (Reinelt et al. 2001).
Monthly or seasonal mean water level	
Graphical analysis of wetland water level and/or water table level dynamics over time	This metric is a plot of water level over time.

like a hydrologic signature" (Mitsch and Gosselink 2000, describing *hydroperiod*). It is difficult to describe a complex and annually variable time series with simple mathematical terms, so a large number of metrics can be used to characterize a wetland's hydropattern (Table 2.9). Nuttle (1997) recommended that hydropattern measurements incorporate the following four distinct attributes of the water-level time series: (1) average water level for a period, (2) intensity or amplitude of fluctuation, (3) cyclic periods embedded in fluctuations, and (4) timing of fluctuations with respect to life histories of interest. Similarly, Wissinger (1999) proposed that "biologically relevant" classifications of wetland hydropatterns should include the following components:

Permanence (permanent, semi-permanent [dry in some years], temporary),

Predictability of drying and filling,

Phenology (seasonal timing) of drying and filling,

Duration of the dry and wet phases, and

Harshness during both phases.

Wetland hydropatterns vary hugely between wetlands of different types and climates and also vary within wetlands of the same type and climate. Wissinger (1999) neatly summarized the many types of hydropatterns observed in North American wetlands (Fig. 2.16), and Mitsch and Gosselink (2000) provided excellent examples of various hydopattern behaviors.

Hydropattern is a dominant driver of wetland ecology. Permanently inundated wetlands usually support fish (and other top predators such as bullfrogs), the presence of which precludes the existence of many amphibian species and alters macroinvertebrate community structure. Conversely, wetlands with very short hydroperiods are not wet long enough for many amphibians to breed and mature beyond their larval stage (e.g., Pechman et al. 1989; Snodgrass et al. 2000; Ryan and Winne 2001). Broad differences in hydropatterns are reflected in different macroinvertebrate community structures (e.g., King et al. 1996; Wissinger et al. 1999; Brooks 2000; Acosta and Perry 2001; Boix et al. 2001), but small differences in hydropatterns do not perceptibly alter macroinvertebrate communities (Batzer

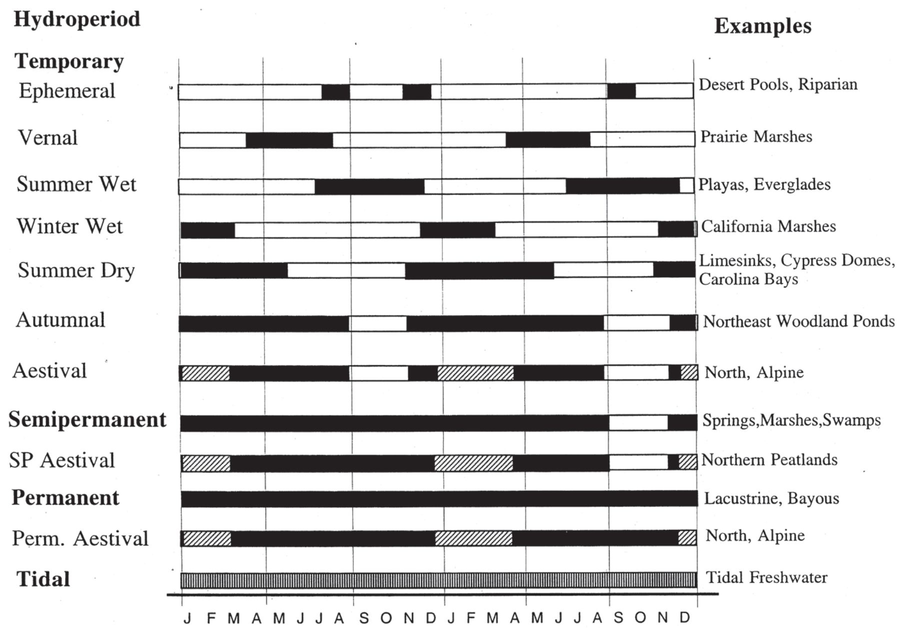

FIGURE 2.16. Summary of hydropatterns observed in North America. Dark blocks represent inundation, hatched blocks represent iced conditions, and white areas represent dry periods. From Wissinger (1999).

et al. 2004). The greatest amphibian and macroinverebrate richness is usually found in wetlands with long, but not perennial, hydroperiods (Whiles and Goldowitz 2001; Paton and Crouch 2002). Hydropatterns affect plant communities not only through length of inundation, but also through depth of inundation (e.g., David 1996; Newman et al. 1996; Baldwin et al. 2001; Steven and Toner 2004). Water levels of some wetlands are managed to create optimal conditions for plants that produce high-quality food for waterfowl. Hydropatterns also affect decomposition rates of organic material (e.g., Battle and Golladay 2001; see also Chapter 4).

Wetland hydropatterns can be observed simply and inexpensively with a staff gauge visited on a regular schedule, but staff gauge observations have several drawbacks. Staff gauges provide no information on water-level dynamics occurring between visits, and human observers have trouble maintaining a precisely regular visitation schedule. Automatic water-level monitoring using any of a variety of water-level sensing and data recording technologies can provide continuous water-level data sampled at short intervals at a higher but reasonable monetary cost. Automatic sampling equipment measurements must be checked regularly for sensor drift, temperature effects on measurements, and power supply status. Equipment failure for any of a number of reasons may leave data gaps unless redundant sampling equipment is installed. Automatic water-level monitoring systems are usually installed in a well and thus provide information on water depth when the wetland is inundated and water table depth when the wetland is dry.

Non-recording wells are inexpensive, easy-to-install, and provide information on shallow water table dynamics when the wetland is not inundated. A small diameter tube cut to the same length as the well can be left inside the well to provide information on the highest water level since the last visit. Finger paint or water-finding paste can be streaked on the tube prior to insertion into the well, and the highest smearing or color change of the paint/paste indicates the previous high-water level. Alternatively, cork dust can be placed in the well, and the cork dust will leave a ring on the inside tube at the highest water level since the previous visit. Morgan and Stolt (2004) compare various approaches for monitoring water table and water-level time series.

Bridgham et al. (1991) developed a steel rod oxidation methodology to quantify relative differences in water table depth in wetland soils. Steel rods are inserted into the ground and left in place for an extended time period. Where the rod is oxidized, water levels have risen and fallen, allowing oxygen into the soil and causing oxidation of the rod. Where the rod is free of rust, it is assumed that the soil has been nearly continuously saturated. This technique allows a semi-quantitative evaluation of relative wetness/dryness of surface soils. Jenkinson and Franzmeier (2006), Rabenhorst and Burch (2006), and Castenson and Rabenhorst (2006) all refined this idea with ferrihydrite coatings on PVC pipe, and achieved better interpretation and more precise delineation of redox conditions.

Wetland Hydraulics and Residence Time

Hydraulics is the study of the physics of water bodies, both moving (*hydrodynamic*) and static (*hydrostatic*). The hydraulics of wetlands affect sediment deposition, biogeochemical cycling, residence time, and wetland hydropattern.

Most wetlands are slow water environments (exceptions include tidal marsh channels and some riverine wetlands), so when surface waters enter a wetland carrying sediment particles, some of the sediment will settle out along with

SIDEBAR 2.3 STOKE'S LAW DESCRIBING THE FALL VELOCITY OF PARTICLES IN SLOW-MOVING LIQUIDS

The fall velocity of a spherical particle in slow-moving liquids is described by Stoke's law:

$$v = (g/18\nu)(\rho_s - \rho)d_p^2$$

v = particle fall velocity in water, cm/s,

g = acceleration of gravity, 981 cm/s^2,

ρ_s = density of sediment, g/cm^3,

ρ = density of water, 1.0 g/ cm^3, and

d_p = particle diameter (cm)

ν = absolute viscosity of water, a function of temperature, cm^2/s.

Stoke's law says that fall velocity increases with the difference between the particle density and the fluid density and also with the square of the particle radius. For example, sand particles are relatively large and settle rapidly in water while clay particles are very small and may remain in suspension for months (Table 2.10).

TABLE 2.10
Settling Velocities of Soil Particles at 10°C

Particle type	Assumed Diameter (mm)	Velocity (mm/s)	Approximate Travel Time per Meter
Coarse sand	1 mm	100 mm/s	10 s/m
Fine sand	0.1 mm	8 mm/s	2 min/m
Silt	0.01 mm	0.154 mm/s	2 hour/m
Clay	0.0001 mm	0.0000154 mm/s	750 days/m

nutrients or other pollutants adsorbed to the sediment particles. The amount and type of sediment that will settle out depends on the size distribution of sediment particles transported to the wetland and the residence time of the wetland. Larger sediment particles fall faster than smaller sediment particles because the mass of the particle increases with the cube of the particle radius, while the friction forces that are acting on a particle moving through a liquid increase with the square of the particle radius (Sidebar 2.3, Table 2.10).

Wetlands receiving streamflow carrying sediments usually display spatial patterns in soil texture. Because sands settle rapidly, soils near the stream discharge point tend to be sandier whereas soils in slack coves of the wetland tend to have fine-textured soils. Over time, deltas form where streams enter wetlands. The amount and type of sediment entering and exiting a wetland strongly controls plant successional processes and the evolution of wetland morphology. Sediment deposition and organic matter accumulation cause shallow lakes to become wetlands and wetlands to become meadows.

The *residence time* of a wetland is the average time a water molecule spends in a wetland from the time it enters to the time it leaves. A short residence time means that water is flushed quickly and undergoes relatively little biogeochemical transformation in the wetland while a long residence time means the opposite. Conceptually, residence time is easy to calculate:

$$R_t = V/Q \tag{2.16}$$

where

R_t = residence time of wetland, seconds,

V = volume of water in wetland, m^3, and

Q = outflow rate of wetland, m^3/s.

A large wetland with relatively small outflow, such as the Okefenokee Swamp in Georgia, will have a very long residence time. Such a wetland will be a sink for pollutants that enter the wetland, such as mercury from atmospheric deposition (George and Batzer 2008). Conversely, a small beaver pond on a relatively large stream will have a very short residence time.

For a given average water depth, more and finer sediment particles will settle in a wetland with longer residence time (there is more time for particles to settle out). For a given residence time, more and finer sediment particles will settle in a wetland with shallower average depth (there is decreased settling distance). In most wetlands with surface waters flowing through them, the residence time is spatially variable, with lower residence times in relatively unobstructed and fast flow paths and longer residence times in backwater areas (Ferguson 1998). Residence time is also temporally variable because neither wetland water volumes nor outflow rates are constant over time.

Biogeochemical processes in wetlands are strongly affected by residence time. Nutrient cycling is dependent on residence time because of the time scales associated with food webs. Sufficient time is needed for the various trophic levels to collect and transform nutrients. If water flows through the system faster than nutrients can be assimilated and cycled, then nutrient sequestration and transformation is not possible. For these reasons, residence time is one of the main design factors in constructed wetlands for stormwater and wastewater treatment.

Hydraulic engineers describe wetlands, ponds, and lakes by their *stage-storage-discharge relationships* (e.g., Dingman 1994; Ferguson 1998), which define the volume of water stored in a wetland and the wetland outflow for any given water depth (stage). Hydraulic relationships are usually monotonic, meaning that for any stage, there is only one possible outflow rate and one possible storage volume, although this assumption breaks down when there are backwater effects of surface waters downstream. The outlet configuration of a wetland strongly affects the stage-storage-discharge relationship. For example, if all other hydrologic factors are equal, a wetland that drains over a long beaver dam will have very different hydropattern characteristics than the same wetland draining through a small culvert. In the case of the beaver dam, a slight rise in stage produces a large increase in outflow rate, so the wetland water level changes very little during a storm. When the wetland drains through a culvert, the stage must rise much higher to achieve the same discharge rate, so the water-level fluctuation during a storm is large. There are infinite possible outlet configurations for wetlands, and each will alter a wetland's hydropattern. Another complicating factor for hydrodynamics of large wetlands and estuaries is wind fetch, which can move water directly by shear stress or by creating hydraulic gradients as water "piles up" on the downwind portion of the wetland. In some cases, wind effects can override the usual forces driving flow dynamics (e.g., Yallop and O'Connell 2000; Moreno-Ostos et al. 2007).

Geomorphic Controls on Wetland Hydrology

Wetlands occur where hydrologic conditions driven by climate, topography, geology, and soils cause surface saturation of sufficient duration to form hydric soils and competitively favor hydrophyllic vegetation (Fig. 2.17). From a geomorphic standpoint, wetlands occur in a relatively limited set of positions (Table 2.11), including in depressions like prairie potholes or bogs, in nondepressional settings such as floodplains, and in high water table areas or groundwater discharge areas, such as on continental margins.

From a geologic perspective, wetlands are constantly being destroyed, shifting locations, and being created. Almost all wetlands are late Pleistocene or Holocene features usually less than 12,000 years old (the authors can find no exception to this general rule, but cannot rule out the possibility of exceptions). The Holocene is the geologic epoch from the end of the last episode of continental glaciation to the present. Some wetlands, such as beaver swamps, have lifespans of just a few decades. Some areas of wetland concentrations, such as the lower Mississippi Valley and the Amazon, are ancient, but the wetland features themselves are constantly shifting because of river and sediment dynamics. The primary drivers of natural wetland destruction and creation are eutrophication, sedimentation, erosion, glaciation, climate change, declines or ascensions in water tables, and sea level change. Obviously, these factors are not independent.

Every wetland is a unique actualization of many abiotic and biotic factors, including geologic and geomorphic history, topography, connections to the local and regional hydrologic system, connections to local and regional ecosystems, time since formation, and disturbance history. For example, although they will share obvious similarities, no two beaver swamps are identical. Paraphrasing Gertrude Stein, "a wetland is not a wetland is not a wetland." Consequently, wetlands defy simple classification. Wetland scientists have tried vegetative and geomorphic classification systems (Cowardin et al. 1979), hydrologic classification systems (e.g., Cowardin et al. 1979; Bridgham et al. 1996; Reinelt et al. 2001), hydrogeomorphic classification systems (Ramsar Convention Bureau 1991; Brinson 1993a), water chemistry and hydrology classification systems (Warner and Rubec 1997), and others. While all these systems have value, none have provided robust classification and description for all wetlands. What, then, is the value of the geomorphic classification presented in Table 2.11? This table

TABLE 2.11
Geomorphic Classification of Major Wetland Types

Wetland Types	Notes and Examples
Stream and River Associated	
Floodplain or alluvial wetlands	Hydrology of floodplain wetlands is affected variously by high water tables, frequent overbank flow, or backwater effects of hydraulic impediments to streamflow. Floodplain wetlands tend to occur in depressions formed by fluvial action. See Figure 2.18.
High floodplain water table	Wetlands occur anywhere the water table lies at or near the floodplain surface for significant duration.
Overflow channels	An example is a backwater swamp.
Abandoned oxbows	Relict river bends that are isolated from current channel
Paleo channel features	Relict low spots in floodplains are created by formerly active flow paths.
Beaver swamps	These swamps are created in small and medium streams.
Accidental swamps	These are created when humans inadvertently inhibit the drainage of a stream with road crossings with insufficient culvert capacity or the placement of railroad grade fills across valleys.
Ill-defined stream and river channels	These are created by low dams or drainage blockages. They are also common where mountain river systems drain into low gradient areas and overbank flow is frequent and of long duration (e.g., Pantanal, Amazon).
Lacustrine	
Lake fringes	Lakes with low water level variation develop wetland fringes.
Lake deltas	Wetlands form where sediment-laden streams enter lakes.
Dying lakes	Small alpine lakes eventually become alpine wetlands and then flat meadows.
Glacial	
Glacial till	Bogs, fens, and tundra are examples of wetlands that form on flat, infiltration-limited topography common in glacial till deposits.
Glacial outwash	Prairie potholes and kettles form where relict blocks of ice left depressions in the glacial outwash. Kettles become wetlands only if regional water tables intersect kettle bottoms for a period of time sufficient for fines and organic matter to accumulate and seal the wetland bottom. Wide flat outwash channels feature a variety of wetland types.
Groundwater associated	
Lower coastal plains	Low-lying flat areas of coastal plains typically feature near-surface water tables caused by the hydraulic control of the ocean, and these areas feature a variety of isolated depressional wetlands (cypress domes and gum swamps); linear, seasonally wet areas (hardwood bottoms); as well as large contiguous wetland areas (pocosins, Tate's Hell Swamp in Florida, Okefenokee Swamp in Georgia).
Hillslope seeps	Geologic discontinuities may force hillslope groundwater and/or interflow to exit the side of the hill and thus create wetland conditions.
Regional discharge areas	Desert oases are examples.
Limestone dissolution areas	These occur over limestone geology in a variety of geomorphic settings including mountain ridgetops. Sinkholes in Florida are well known examples, as are Carolina Bays.

(continued)

TABLE 2.11. *(continued)*

Marine associated	
Fresh water marshes	Marshes that occur along rivers and distributary channels as rivers enter estuarine environments. The hydropatterns of these marshes are affected by tides, but the principal source of water is fresh river water.
Saltwater marshes	Marshes with tidal hydropatterns and high salinities caused by the mixing of seawater and river water. Salinities in these environments occur on a continuum.
Estuarine fringe	Shrubby or forested wetland areas are on the upland fringes of estuarine marshes.
Miscellaneous inland depressions	
Wind-carved depressions	These depressions are most common in arid areas. Some may form wetlands such as playas.
Paleo-marine features	A variety of inland depressions are left by ancient marine processes. Examples include California's vernal pools.
Manmade depressions	Old quarries and borrow pits commonly form wetlands after sufficient time for hydric soil development and hydrophyllic plant colonization.
Animal-made depressions	Examples include buffalo wallows.

NOTE: Figures 2.17 and 2.18 provide some examples of wetlands associated with streams, groundwater, lakes, and oceans.

illustrates that, regardless of the biotic and structural diversity found in the world's wetlands, there are only a few general landscape locations in which wetlands occur and a few processes by which wetlands are formed. Such a geomorphic understanding of wetlands helps explain and describe the spatial and temporal context of wetlands within the biogeography of a region. Furthermore, a simple geomorphic view of wetlands is instructive for setting regional and national wetland management policies and for designing created wetlands. With the exception of the polar environments, wetlands occur all over the world, on all continents, and in all climates, but almost all occur in the few landscape positions listed in Table 2.11.

Wetlands are relatively abundant on active floodplains because the water table draining to the stream/river is near the floodplain surface and because overbank flows from the stream/river may cause periodic inundation. Floodplains are dynamic and complex features that, over time, collect a variety of fluvially scoured topographic features (e.g., Hupp 2000; Leigh et al. 2004), some of which are conducive to wetland formation. Floodplain wetland features can be transformed immediately by large floods or other natural disturbances (e.g., Johnson et al. 2000). Figure 2.18 illustrates the variety of wetland and fluvial features that may be found on a large river floodplain. Depending on its geomorphic history, a floodplain may have several levels, including an active floodplain that receives overbank flows during floods and several terraces comprising the paleo floodplain. Terraces are usually relatively high above the water table and rarely, if ever, receive overbank flow. Thus wetlands tend to occur on active floodplain areas, and less so on terraces.

Natural lakes with muted water-level fluctuations have wetlands on low-slope lake margins, in coves, and on sediment deltas where streams and rivers discharge. Conversely, the management of artificial reservoirs for water supply and flood control usually results in large annual water-level fluctuations that do not favor wetland formation on reservoir margins. Reservoirs, therefore, usually feature wetlands only in delta areas. A comparison of lacustrine wetlands between natural and artificial lakes illustrates how water-level dynamics can control wetland formation and characteristics.

Continental glaciers are powerful agents of wetland creation. Glacial till, compacted by the weight of the overlying ice, tends to have very low hydraulic conductivity. In locations where precipitation exceeds evapotranspiration for much of the year, relatively flat till deposits form wetlands in even slight depressions because of the restricted percolation rates of till soils. Outwash soils—typically sands and gravels discharged from a glacier but not compacted by it—have high hydraulic conductivities, which would suggest that these soils are unlikely places for wetland formation unless they intersect the water table. Flat outwash channels, however, often act as conduits for regional groundwater flow, and water tables lie at or near the ground surface. Therefore outwash channel features are often wetland areas. In addition, kettles or potholes left by blocks of ice calved from retreating glaciers often become wetlands. The Upper Midwest and Northeast of the United States features numerous small and large wetlands created on glacial till and outwash features, many of which are organic soil wetlands, or peatlands.

Many if not most wetlands are hybrids in terms of their

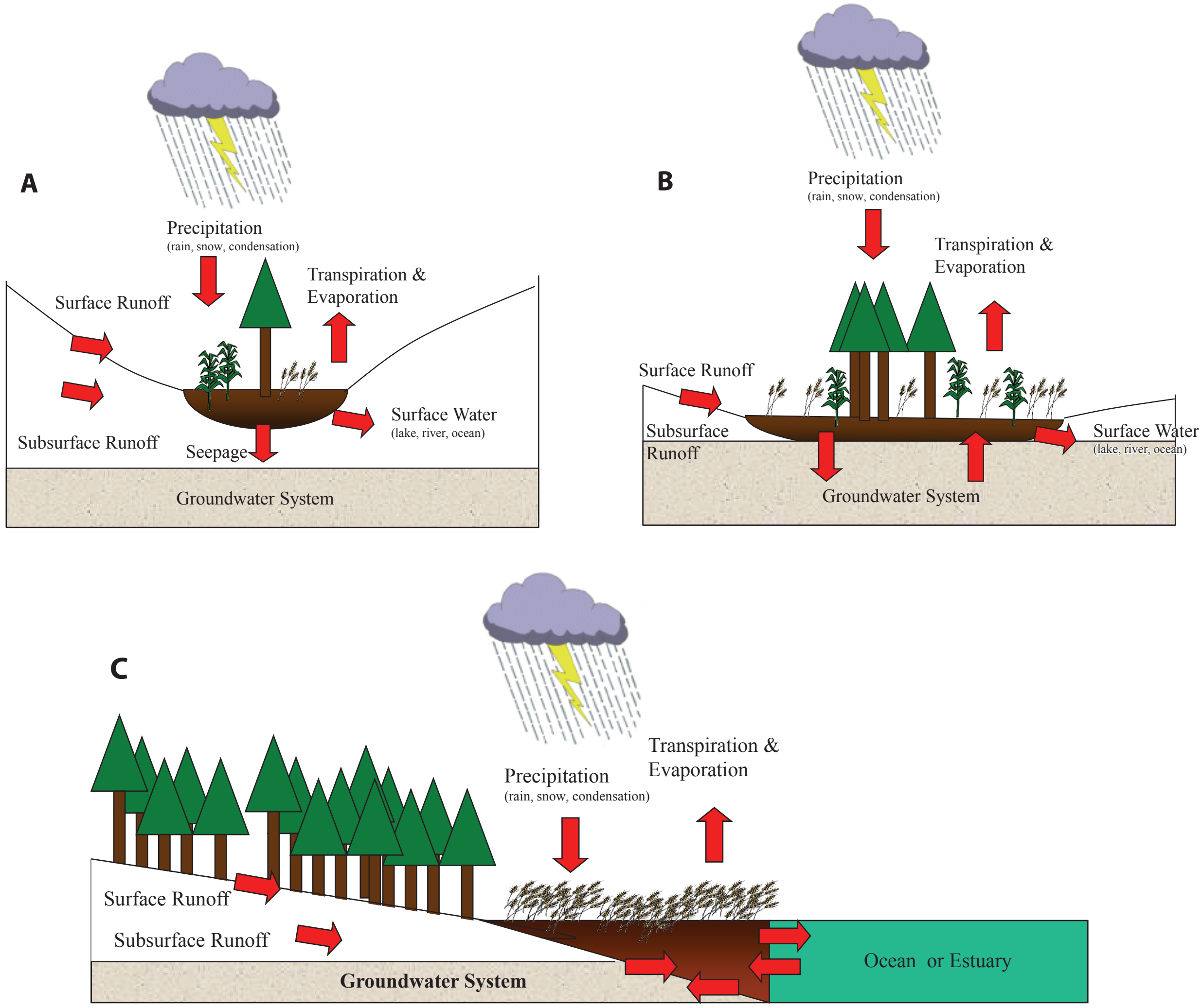

FIGURE 2.17. Typical wetland geomorphic positions including (A) depressional, (B) non-depressional, and (C) tidal or estuarine.

geomorphic origins and controls. Floodplains contain many wetlands partly because of shallow water tables, so floodplain wetlands could be considered groundwater-associated wetlands. Furthermore, groundwater-associated wetlands of coastal plains are influenced by sea-level controls on groundwater dynamics, so they could be considered marine associated. Still, it is difficult to think of a wetland that does not fit into one or more of these simple geomorphic categories.

Some wetlands, such as the Carolina bays of the eastern seaboard of the United States, are difficult to classify. Carolina bays are carbonate dissolution features that occur on marine deposits in both the upper and lower coastal plains overlying geologic units containing carbonate. Many Carolina bays are perched well above the normal surficial water table and located far outside of floodplains and the area of marine influences on groundwater. Carolina bays could be considered to be groundwater associated (because of the groundwater processes that caused the carbonate dissolution), or they might be limestone-dissolution areas, or they may be paleomarine features (given their landscape position).

The distinction between marine-associated wetlands and many wetlands of groundwater origin is blurry. The marine environment is the ultimate discharge point for aquifer systems, and sea level provides the base hydraulic control for aquifer discharge. As a result, many low-gradient coastal plain areas feature very shallow water tables, and thus wetlands are common in topographic depression in coastal plains. Hydrologic inputs to coastal plain wetlands are precipitation and possibly surface flow and groundwater inputs, but the dominant hydrologic control on coastal plain wetlands are seasonally high water tables. In the continental interior, groundwater-dominated wetlands occur where large regional groundwater systems discharge or connect to flat valley bottoms.

Marine-associated freshwater marshes receive most of their water from river flow that has backed up and spread out because of the hydraulic control exerted by the tidal

TABLE 2.12
Benefits of Forested Buffers around Wetlands and Considerations of Width and Function

Function	Necessary Width	Notes
Sediment, nutrient filtration	Width varies depending on upslope conditions. More is better. 10m – 30m	This is most important during construction to minimize sediment input during excavation, grading, and grass establishment. It also helps keep sediment, fertilizers, and pesticides from upslope activities such as agriculture and silviculture out of wetlands.
Minimization of direct human impact	15m is usually adequate.	People living near wetlands may landscape down to the edge or dump yardwaste and garbage in the buffer or the wetland.
Organic debris input	Narrow buffer (~3m) will suffice.	Leaves, twigs, pine needles, and so on can be important drivers of wetland ecology.
Large woody debris recruitment	½ mature tree height (12m – 20m)	Pieces of large wood provide habitat complexity, substrate for macroinvertebrates, cover for amphibians, roosts for waterfowl, and haul-out and sunning areas for turtles.
Wildlife habitat	Benefits increase out to 100m.	Many birds and amphibians depend on terrestrial habitat adjacent to wetlands. It is an important remnant habitat after basin development.
Bank stability	3 m will suffice.	Bank stability is sometimes valuable on certain wetland features with well-defined banks (e.g., sand rims on Carolina Bays, streambanks on floodplain wetlands).
Shade (water temp)	15 m is generally sufficient.	Shade is more important for forested wetlands where light limits primary productivity. Peripheral shade is unimportant for marshes and open water wetlands.
Social (children playing)	>12m.	A narrow buffer grows a lot of invasive plants and blackberries and briers due to light effects. It is hard to play in a narrow buffer.
Aesthetics	Eye of the beholder	After development, wetland and stream buffers are often the only natural areas left in the landscape.

systems into which they discharge. Water exchange in salt water marshes is predominantly by tidal exchange. In between, brackish marshes receive both river and tidal inputs.

Wetlands are a product of a landscape's geologic and geomorphic history as well as its current hydrologic behavior. A landscape's hydrologic behavior is driven by topography, soils, climate, vegetation, and land use. Once formed, wetlands can modify local or regional hydrology. For example, depressional flow-through wetlands, such as beaver ponds, temporarily store flood flows and reduce peak flows downstream. They also trap sediment, thus altering stream dynamics.

There have been many efforts to develop systematic evaluations of the hydrologic functions and values of wetlands (e.g., Adamus and Stockwell 1983; Bardecki 1984; Carter 1986; Hruby et al. 1995; Smith et al. 1995; Brinson and Rheinhardt 1998), but the hydrologic functions of wetlands have in some cases been overstated. Floodplain wetlands are often cited as providing floodwater storage and flood mitigation, but any part of the floodplain does the same thing, and this is not a function peculiar to jurisdictional wetlands. Bullock and Acreman (2003, p. 368) extensively reviewed the literature on hydrologic functions of wetlands and found "generalised and simplified statements of wetland function are discouraged because they demonstrably have little practical value." They concluded that "apparently similar wetlands are driven by very different hydrological processes; almost invariably, some data need to be collected at a site to identify its functional role."

Effects of Land Use on Wetland Hydrology

Land use activities adjacent to wetlands can affect wetland habitat by altering inputs of sunlight, sediment, organic debris, nutrients, dissolved carbon, and sometimes contaminants such as pesticides, heavy metals, and organic chemicals. Vegetation and soil alteration in a wetland's watershed predominantly affects the wetland through altered hydrology and sediment contributions. Some activities

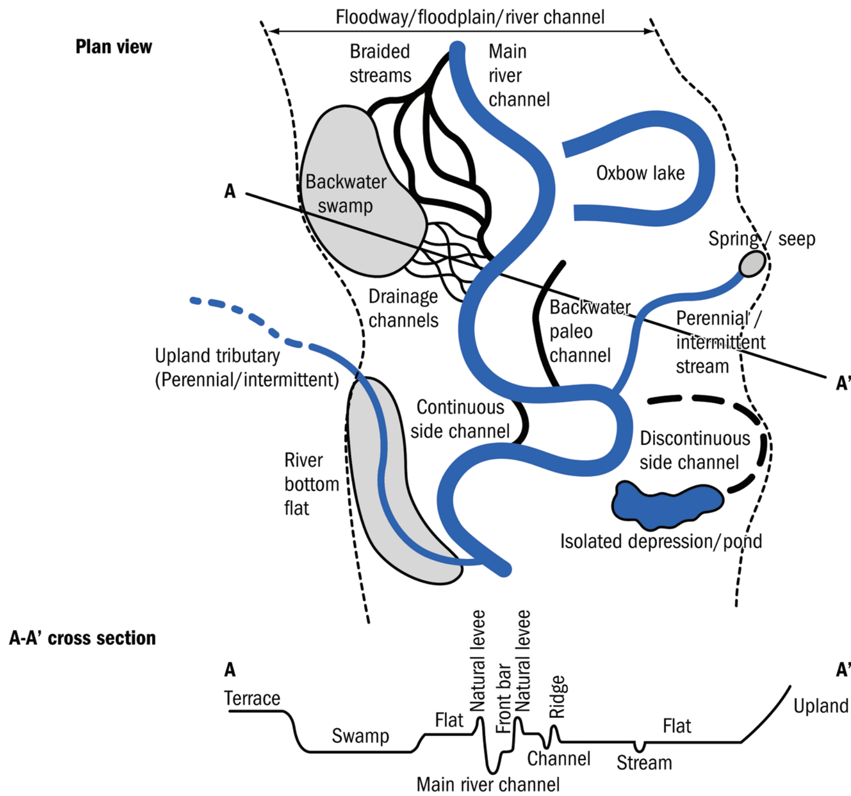

FIGURE 2.18. Varieties of floodplain wetland features. Figure courtesy of Masato Miwa, Kyushu University. Modified from Mitsch and Gosselink (2000) and Hodges (1998).

may directly alter outlet hydraulics and thus change a wetland's hydropattern. Depending on the details of the activity, abiotic effects of human actions may make a wetland wetter, drier, flashier, sunnier, or more nutrient-rich, and these abiotic changes may alter the biology of the system. Basic hydrologic concepts can be applied to predict likely effects of landscape alteration around and above a wetland. A body of literature has examined effects of forestry, agriculture, and urbanization on wetlands in various settings.

Clearcutting a wetland's watershed reduces interception and evapotranspiration sufficiently to cause water table rise for one to five years following harvest (Riekerk 1985; Riekerk 1989a, 1989b; Aust and Lea 1992; Crownover et al. 1995; Dube et al. 1995; Lockaby et al. 1997; Sun et al. 2000; Bliss and Comerford 2002). This water table rise increases inundation periods and reduces dry periods in wetlands in the affected basins. Compared to typical rotation lengths for commercial timber (as little as 22 years in the southeastern United States and as much as 70 years in the Pacific Northwest) the hydropattern effects of canopy removal are relatively short-lived. If the canopy of a forested wetland is cut without removing the canopy in the contributing basin, an opposite hydropattern effect may occur because direct evaporation from the wetland will increase.

Observed effects of this hydropattern change on plants and macroinvertebrates are mixed, because the hydrologic effect is difficult to divorce from light effects or nutrient effects of adjacent forestry practices. For example, Batzer et al. (2000) studied wetlands in commercial pine plantations and found that wetlands surrounded by smaller trees had greater light levels, water temperatures, pH, herbaceous plant cover and biomass, terrestrial invertebrate diversities and numbers, and water flea numbers. These wetlands also had lower specific conductivities and aquatic oligochaete numbers than wetlands surrounded by mature trees. Hydrologic differences were not detectable, and it was hypothesized that these differences were driven by differences in nutrient concentrations and light. Even without fertilization of newly planted trees, timber harvest usually results in a temporary release of nutrients, principally nitrogen (Riekerk 1985; Dahlgren and Driscoll 1994).

The effects of agricultural practices on nearby wetlands are difficult to generalize. Irrigation of agricultural lands may increase or decrease inundation periods of adjacent wetlands, depending on whether irrigation water is pumped from the surficial aquifer, a confined aquifer, or a nearby stream (Bolen et al. 1989; Smith and Haukos 2002). Tilled fields increase surface runoff and thus increase water-level fluctuations in receiving wetlands (Euliss and Mushet 1996). Tilled fields also increase sedimentation in receiving wetlands (Martin and Hartmann 1987; Luo et al. 1997, 1999).

Azous and Horner (2001) provide many case studies from the Puget Sound basin of the Pacific Northwest, United States, on the effects of urbanization on wetlands. In general, urbanization increases water-level fluctuations and nutrient concentrations of wetlands. The biotic response to these abiotic drivers is usually reduced plant and amphibian species richness and diversity. In many urbanized wetlands of the Puget Sound, hardy invasive plant species outcompete native vegetation.

Undisturbed vegetative buffers are required or recommended by many state and local laws and by some forestry and agricultural best management practice manuals to mitigate the effects of land use change on wetlands. Vegetative buffers provide a number of benefits to wetlands (Table 2.12). Federal law provides some protection against direct alterations of wetlands, but federal laws provide no buffer protections. Horner et al. (2001) provide guidelines for maintaining wetland functions in developing basins, and Booth et al. (2002) provide guidelines for protecting the natural hydrology of developing basins. All such guidelines include buffers as a necessary protection for wetlands. The buffers themselves provide little to no protection against hydrologic change, but they minimize the other abiotic influences associated with human land use activities.

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Climate, geology, topography, and soils interact to drive the hydrologic pathways and fluxes in the landscape, and at larger scales they control the spatial distributions of wetlands in our landscape (Fig. 2.1). For each wetland, the hydrologic fluxes and hydraulic controls determine the hydropattern that in turn sets the template for soil development and vegetative dynamics. Wetland ecosystem functions, such as processing carbon and nutrients, transforming, or mineralizing pollutants, providing habitat, and metering floodwaters and sediment pulses are dependent on the hydrologic setting.

Soils that are permanently or periodically saturated by slow-moving water develop characteristics fundamentally different from upland soils. Slow water allows fine sediments and organic detritus to settle, and the lack of oxygen changes redox conditions, alters the mobility of minerals, and slows down decomposition rates. The resulting hydric soils create plant and microbial habitats quite different from those found in uplands or deep waters. A basic understanding of the hydrologic cycle, geomorphic processes, and soil development allows the development of conceptual models of how wetlands form and function anywhere on the planet.