Chapter 5

Wetland Hydrology

The purpose of this chapter is to describe and discuss the general hydrologic properties that make wetlands unique, and to provide an overview of the processes that control wetland hydrologic behavior. The intent is to provide a general discussion of wetland hydrologic processes and methods in the hope of fostering an understanding of the important attributes of wetland hydrology relevant to the monitoring and assessment of these systems.

As such, it is not intended to address the narrower definition of wetland hydrology for jurisdictional or classification purposes. Also, this module should not replace more advanced wetland texts. If the need arises to obtain more specific information, the reader is advised to refer to wetland books or articles, including those referenced within this document.

5.1 Introduction

Wetlands are a unique hydrologic feature of the landscape. One particularly important attribute is their position as the transition zone between aquatic and terrestrial ecosystems. Wetlands share aspects of both aquatic and terrestrial environments because of this position. On one hand, most freshwater and marine aquatic environments, such as lakes, rivers, estuaries, and oceans, are characterized as having permanent water. On the other hand, terrestrial environments are generally characterized as having drier conditions, with an unsaturated (vadose) zone present for most of the annual cycle. Wetlands thus occupy the transition zone between predominantly wet and dry environments.

A diagnostic feature of wetlands is the proximity of the water surface (or water table below the surface) relative to the ground surface. In freshwater and marine aquatic habitats, the water surface lies well above the land surface, while in terrestrial environments it lies some distance below the root zone as a water table or zone of saturation. The shallow hydrologic environment of wetlands creates unique biogeochemical conditions that distinguishes it from aquatic and terrestrial environments.

Geomorphic Position

Wetlands are a fundamental hydrologic landscape unit (Winter, 2001) that generally form on flat areas, or on shallow slopes, where perennial water lies at or near the land surface, either above or below. Wetlands tend to form where surface water and ground water accumulate within topographic depressions, such as along flood plains, within kettles, potholes, bogs, fens, lime sinks, pocosins, Carolina Bays, vernal pools, ciénegas, pantanos, tenajas, and playas, and behind dunes, levees, and glacial moraines. Seepage wetlands form where ground-water discharges on slopes, as well as near the shore of streams, lakes, and oceans. Fringe wetlands also form along shorelines, with periodic inundation not caused by ground-water discharges, but rather by water exchanges with adjacent waterbodies, such as by periodic floods and tidal action. And finally, perched wetlands form above low-permeability substrates where infiltration is restricted, such as above permafrost, clay, or rock (Novitzki, 1989).

Brinson (1993) provides a methodology for using hydrogeomorphic indicators to classify wetlands based on their unique hydrologic, geomorphic, and hydrodynamic characteristics. In this way, the dominant landscape and hydrologic factors can be synthesized to better develop an understanding of wetland forms and functions.

Energy as the Driving Force

The direction and rate of water movement into and out of wetlands is controlled by the spatial and temporal variability of energy. A change in energy with distance generates a force that causes water to move from zones of high energy to zones of lower energy. Gravitational forces account for most water movement, in that water tends to flow from higher to lower elevations. Resisting the gravitational force are viscous (friction) forces that retard the fluid velocity. Inertial (momentum) forces resist a change in velocity, causing water to move at a constant velocity, and in a straight line, unless additional energy is expended to either accelerate, decelerate, or deflect the water.

Water can also move due to a change in pressure, from zones of high pressure to zones of low pressure. This is common in ground-water systems, where confined aquifers flow to the surface because of the greater pressure at depth. Artesian flow from a confined aquifer to the surface occurs when the recharge area to the aquifer lies at a higher elevation than the ground surface where the discharge occurs. Classical artesian springs exist in low-lying areas that are supplied with flows from higher elevation areas.

Wetlands are normally found in low-energy environments - that is, in areas where water normally flows with a slow velocity. This results, in part, because the land surface is relatively flat in these areas (Orme, 1990). Because wetlands lie in relatively flat landscapes, their surface area expands and contracts as the water stage changes, allowing for the storage of large volumes of water. Wetlands therefore serve as a moderator of hydrologic variability - storing flood flows and reducing flow velocities during wet weather in particular. In addition, shallow depths and low slopes, consistent with low energy environments, are important for trapping nutrients and sediments.

5.2 Hydrologic Measures

Three hydrologic variables can be defined that are useful for characterizing wetland hydrologic behavior: the water level; the hydropattern; and the residence time. Each of these wetland descriptors are described in greater detail in subsequent sections. What follows here is a brief introduction of these concepts.

One hydrologic descriptor is the general elevation of wetland water levels relative to the soil surface. Open water usually occurs in deeper areas with few, if any, emergent macrophytes. Any vegetation present in these areas is usually not attached to the wetland bottom, but vegetation may be floating on the water surface. An emergent zone may also be present in areas shallower than the open water zone, containing substantial quantities of emergent macrophytic vegetation, either living or dead. Yet other wetlands may have large areas of exposed, saturated soil that is generally covered with macrophytic vegetation. The water level can, therefore, be used as an indicator of the vegetation types likely to occur in each of these zones.

A second descriptor of wetland hydrology is the temporal variability of water levels. The timing, duration, and distribution of wetland water levels is commonly referred to as the wetland hydropattern, and incorporates the duration and frequency of water level perturbations. The hydropattern of some systems such as tidal marshes, fluctuate dramatically over short periods of time; other systems, such as seasonally flooded bottomland hardwood communities, fluctuate more slowly over time. Yet other wetland systems are more static, and may not display substantial short- or long-term variability. The wetland hydropattern is a function of the net difference between inflows and outflows from the atmosphere, ground water, and surface water.

A third descriptor of wetland hydrology is the residence, or travel, time of water movement through the wetland. Some wetland systems exchange water quickly, having a short duration of time that water remains within the wetland, while water may travel very slowly through other wetland systems. The residence time is the ratio of the volume of water within the wetland to the rate of flow through the wetland. Short residence times occur when the flow through the wetland is large compared to its volume - longer residence times occur when the flow is small compared to its volume. The residence time of a wetland is often related to its hydropattern, in that wetlands with large water level fluctuations may have shorter residence times, such as in tidal marshes. On the other hand, some wetlands may fluctuate rapidly due to large changes in inflow, yet have very long residence times due to slow loss rates.

Wetland Water Level

An important feature of wetlands is the condition of oxygen deficiency in wetland soils. Anaerobic conditions develop more quickly in saturated soils than in unsaturated soils due to low oxygen solubility in water, slow rates of water advection, and slow diffusion rates of oxygen through water. Anaerobic conditions in wetland soils affect vegetation by creating adverse conditions for root survival and growth. Thus, the presence of water substantially affects soil oxygen concentrations, which affects plant growth and survival.

Yet, despite these low oxygen concentrations, wetlands are among the most biologically productive ecosystems on the landscape. They support a diverse assemblage of vegetative species having special physiological adaptations enabling them to survive and prosper in these otherwise harsh growing conditions. Many biogeochemical reactions occur within these low oxygen zones, as noted elsewhere in this document.

Water levels in wetlands serve as an indicator of the dissolved oxygen state of the soil-water system. Wetter systems generally have higher water levels and lower soil dissolved oxygen concentrations, while drier systems have lower water levels and higher dissolved oxygen concentration. A general relationship between wetland water levels and hydric states is shown in Table 5.1 (Cowardin et al., 1979). Note that a distinction is made between soil saturation and soil surface inundation. Some systems may be flooded for part of the year and still have low pore water soil saturation, and vice versa. Low soil dissolved oxygen concentrations and reducing conditions may result in both cases.

Wetland water levels (also called the *stage*) may not be indicative of soil saturation. The zone of pore saturation may extend above the water table due to capillary rise in fine-grained materials (Black, 1996). Capillary rise results from the natural tendency of water to adhere to soil surfaces and to other water molecules. Because the capillary height of rise can extend for several meters above

Table 5.1: Wetland Types

TIDAL WETLANDS:

Subtidal - Tidal water permanently covers the land surface.

Irregularly Exposed - Tidal water usually covers the land surface, but is not exposed daily.

Regularly Flooded - Tidal water alternately covers and daily exposes the land surface.

Irregularly Flooded - Tidal water covers the land surface less oven than daily.

NON-TIDAL WETLANDS:

Permanently Flooded - Water covers the land surface throughout the year in all years. Vegetation is composed of obligate hydrophytes.

Intermittently Exposed - Water covers the land surface throughout the year except in years of extreme drought.

Semipermanently Flooded - Water covers the land surface throughout the growing season in most years. The water table is at or very near the surface when the land surface is exposed.

Seasonally Flooded - Water covers the land surface for extended periods especially early in the growing season, but is absent by the end of the seas in most years. The water table is at or near the surface when the land surface is exposed. Saturated Water never covers the land surface, but the soil is saturated to the surface for extended periods during the growing season.

Temporarily Flooded - Water covers the land surface for brief periods during the growing season, but the water table usually lies well below the surface for most of the season. Plants that grow both in uplands and wetlands are present.

Intermittently Flooded - Water covers the land surface for variable periods with no detectible seasonal periodicity. Long periods of time separate periods of inundation. The dominant plant communities under this regime may change as soil moisture conditions change. Some areas may not exhibit hydric soils or support hydrophytes.

Artificially Flooded - The amount and duration of flooding is controlled by means of pumps or siphons in combination with dikes or dams.

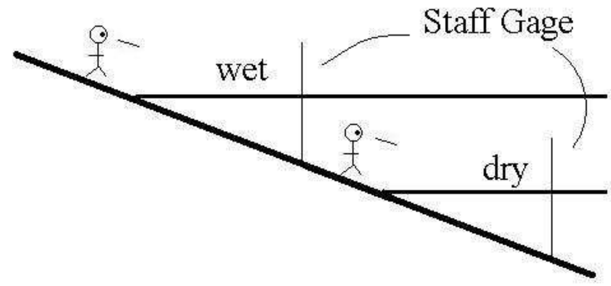

Figure 5.1: Multiple staff gages can be used to determine wetland water levels for a wetland with highly variable stages. The higher gage is used during wet periods, and is easier to read than the more distant one. Once water levels fall below the higher gage, then the lower gage is used.

the water table in fine-grained materials, the soil may be entirely saturated even when water levels are below the ground surface.

Hydrographs A *hydrograph* relates the *stage*, or water level, as a function of time. Between storms, water levels in wetlands normally decline slowly over time, rising in response to precipitation. The *rising limb* of the hydrograph corresponds to the period of time from when water levels begin to rise following a precipitation event. The *peak stage*, corresponds to the time when water levels reach their highest level. The *falling limb* of the hydrograph corresponds to the period following the peak and lasts until the next storm.

The *time to peak* is the length of time between the peak precipitation and peak stage. Times to peak are short in urban areas with large impervious surfaces and channels that have been modified to increase stream velocities. Times to peak are longer in forested areas with few impervious surfaces and channels with many obstructions that slow the passage of water. Another term, the *time of concentration*, is the time required for flow to travel from the most distant point on the watershed, and is a function of the same factors that affect the time to peak.

Monitoring Water Levels Water levels in wetlands can be determined using a staff gage if the water level is above the ground surface. A staff gage is a vertical scale that serves to indicate the elevation of water, or stage, with respect to a reference elevation (Williams et al., 1996). The staff gage is an inexpensive tool that should placed in the wetland such that the base of the staff gage is always submersed in water. For ease of measurement, multiple staff gages can be placed at different depths, such that the nearest one is visible from the shoreline during wet weather. The deeper staff gage is used once water levels fall below the nearer gage (Figure 5.1). In this way, one submerged staff gage is always visible from the shoreline as water levels rise and fall.

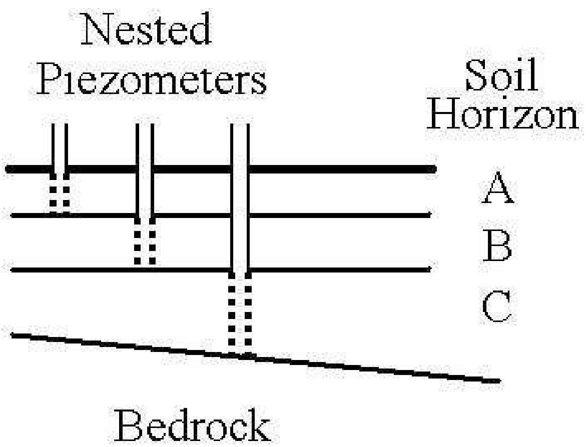

Figure 5.2: Piezometers are used to monitor water level changes in the subsurface. Multiple piezometers (called a *nest*) can be installed next to each other, one in each soil horizon, if vertical flow between soil horizons is expected. Additional piezometers can be installed in the bedrock if such materials affect the hydrology of the wetland.

In some wetlands, a water line can be observed on periodically submerged vegetation. The water line can indicate a high-water level after a flood. In these cases, floating debris - such as leaves, trash, or branches - is lodged in the canopy of trees or bushes. In other cases, the natural water level can be observed as a horizontal line on the sides of trees. This line typically represents the normal water level in the wetland. This line would not be visible during wet weather, but is more likely to be observed during drier periods.

If water levels are below the ground surface, then *piezometers* can be used to find the water surface (Figure 5.2). A piezometer is a small-diameter perforated tube that is installed within the soil at a specified depth (Black, 1996). The perforated zone should be narrow to minimize interference between layers, and placed within a unique hydrogeologic unit, such as a soil horizon or geologic layer.

Water levels can be inexpensively determined by lowering a weighted, chalk-covered steel measuring tape into the piezometer. The tape is lowered until at least one part of the tape is wet. The reading on the tape where the chalk has been wetted is subtracted from the reading taken on the tape at the top of the piezometer. A slightly more expensive technique is to use a depth-to-water detector which provides an audio or visual signal when the water level is encountered. Another option is to use an automated water level recorder, such as a float or pressure transducer. The advantage of automated techniques is their suitability for conditions when water levels are both above and below the ground surface (Black, 1996).

When water levels are below the ground surface, the degree of soil saturation can be measured using Time-Domain Reflectometry (TDR). TDR determines the water content using the electromagnetic properties of a wave pulse passing through a conducting set of rods (such as 3-mm stainless-steel welding rods) placed in the soil (Figure 5.3). TDRs provide the soil water content without the need for calibration. Because the air porosity is not measured using TDRs, additional measurements of the total porosity are required, generally by collecting soil cores. The soil saturation is the ratio of the water content to the total porosity.

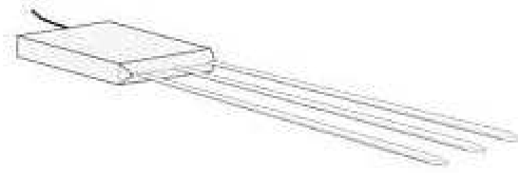

Figure 5.3: Time Domain Reflectometer probes can be used to monitor soil moisture saturation over time, providing an estimate of the water change in the subsurface.

The principle of TDRs is that the velocity of electromagnetic wave along a conductor is a function of the dielectric coefficient of the media around the conductor (Topp, 1980). Larger dielectric constant cause slower wave velocities, and, hence, longer travel times. Liquid water has a dielectric constant of 80.2 at 20°C, while ice is 3.2, petroleum is 1.8 to 2.2, quartz is 4.3, and air is only 1.00. It is clear, therefore, that the wave velocity is substantially retarded as the water content of a soil increases.

Hydropattern

The temporal variability of water levels in wetlands results from dynamic changes in hydrologic inputs and outputs, and temporal changes associated with hydraulic controls within the wetland. Temporal changes in water level are important determinants for many aquatic flora and fauna. The reproductive success of these wetland species can be adversely affected when fluctuations are not correctly synchronized with their developmental stages.

The *hydropattern* is a distinctive feature of the hydrologic variability that describes the variation of water levels over time and space (Acosta and Perry, 2001; King et al., 2004). Hydropattern is a recent term that is used to expand the traditional concept of *hydroperiod* (i.e., the frequency and duration of time that the wetland is saturated) by incorporating additional information about the aerial extent and timing of inundation. The aerial extent is important, especially for large, complex wetlands that contain a variety of wetland features.

Several approaches can be used to characterize temporal changes in wetland stage (i.e., water levels). The easiest approach is to plot wetland stage as a function of time (called the *hydrograph*). The hydrograph shows the stage for a period of time that captures the range of

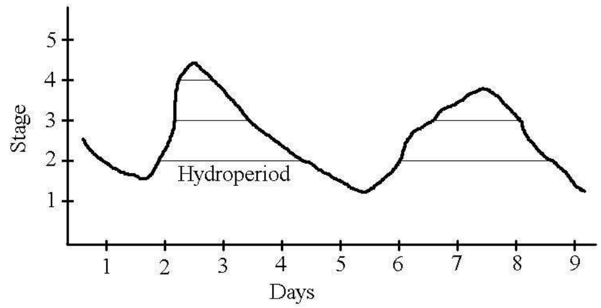

Figure 5.4: Hydrograph for short period of time showing the water level variation. Note that the hydroperiod is marked for a few stages.

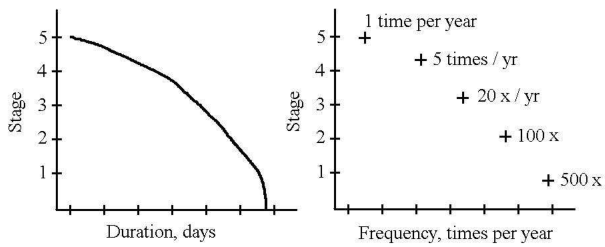

Figure 5.5: Left: Wetland hydroperiod plot showing duration of flooding versus stage. Right: Wetland stage-frequency plot showing number of exceedances per year. Note that the longest duration of flooding occurs at the lower stages, and vice versa. Lower stages have higher frequency of being flooded than higher stages.

possible hydrologic variability (Figure 5.4). Inter-annual, seasonal, event, and daily water level fluctuations may become apparent using such an approach.

Using the observed water levels, a plot of flooding duration versus wetland stage can be constructed (Figure 5.5, left). This plot provides a descriptive summary that indicates how long a typical flood occurs for each stage. Lower elevations have longer durations of flooding than higher elevations. This approach is useful for characterizing water level variability by generating a stage-duration relationship that quantifies the duration in time that a specified water level is exceeded. In this case the period of time that water levels exceed the specified stage (or range in stages) is described. This approach should also consider the seasonal nature of inundations by dividing the data into specific time frames (Mitsch and Gosselink, 2000).

While the stage-duration approach successfully captures the duration of time that the system is flooded, it fails to characterize the frequency with which this occurs. That is, the number of times that a water level exceeds a specified stage for a specified period of time is not quantified. An alternative approach is to quantify the frequency in time that the wetland is observed to exceed a range of specified stages. This approach yields a cumulative frequency table or plot that can be used to calculate ex-

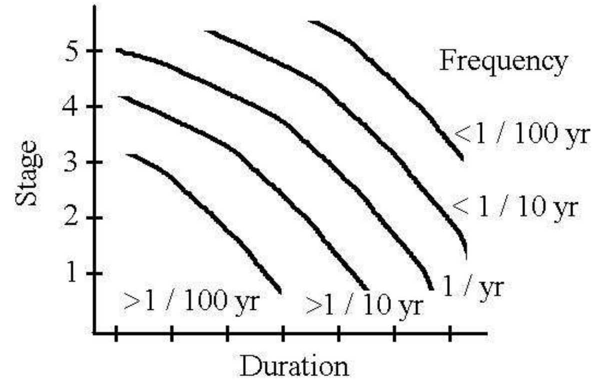

Figure 5.6: Wetland stage-frequency-duration plot showing duration of flooding versus stage for a range of frequencies.

ceedance probabilities (Figure 5.5, right). The mean, median, and extreme stages (e.g., 1, 10, 50, 90, and 99 percentile probability) can be estimated using the exceedance probability plot (McCuen, 1998).

A significant drawback using the exceedance probability approach is that the correlation structure between individual observations may or may not be captured. That is, a system whose water levels vary slowly over time can display the same frequency distribution of water levels as does a system that varies quickly. In effect, the amplitudes of the fluctuations are described, but not the duration. An additional problem is that the frequency diagram, in aggregate, may poorly convey daily and seasonal behavior. Partitioning or stratifying data sets into seasonal or other periods may improve the characterization of water level conditions (Mitsch and Gosselink, 2000).

To overcome these limitations, a stage-duration-frequency, or SDF, curve can be constructed (Figure 5.6). The SDF plot is analogous to intensity-duration-frequency (IDF) curves used in precipitation analysis (McCuen, 1998). SDF curves indicates the frequency that a depth-duration relationship is observed.

For large, complex wetlands, the hydrodynamic behavior may be very different from one area to the next. Characterizing the hydropattern for the entire wetland is a substantially more difficult exercise than for a small, uniform wetland.

Hydrologic Residence Time

The hydrologic residence time is used to evaluate the time required for a hydrologic input to pass through the wetland. The residence time, τ, for a system with constant volume and constant flow rate is simply the ratio of the volume of water within the wetland, V, to the flow rate, Q, or:

$$\tau = \frac{V}{Q} \tag{5.1}$$

The estimated residence time is only appropriate for conditions of 1) piston-flow, such as a First-In-First-Out (FIFO) queue, 2) steady (i.e., constant) flow, 3) single locations of inflow and outflow, and 4) no atmospheric or groundwater exchanges (Himmelblau and Bischoff, 1968). The estimated residence time is not appropriate for conditions when water within the wetland mixes, when multiple inflows and/or exchanges occur at different points within the wetland, or when flow into the wetland is not constant over time.

If these conditions are not met, then the above equation only provides an estimate of the *average* residence time - actual residence times now varying over time and space. Functions for describing the distribution of residence times may be found for simple systems. For example, the exponential function can be used to determine the residence time distribution for a fully mixed system with constant inputs over time (Law and Kelton, 1991).

Different parts of a wetland may exhibit different hydrologic residence times. Water in active, flow-through sections of a wetland may have shorter residence times than water in inactive, isolated parts of a wetland. While each section may have identical hydropatterns, the flow is concentrated in one area, leaving other areas with stagnant conditions. The same equation can be applied regardless, in this case each section would be characterized using the volume of water present in the section, and the flow rate would be characterized using the flow into the section of interest.

Residence times for dynamic systems are more difficult to calculate than steady flow systems. In these cases, the residence time changes - with increasing residence times during periods when outflows exceed inflows, and decreasing when inflows exceed outflows. This is because adding so-called *new* water or removing *old* water from the system decreases the age.

While hydraulic residence times can be calculated using the above equation, tracer tests can also be conducted to confirm these calculations. Conservative tracers (i.e., non-reactive tracers that move passively with the water velocity) can be added at the inflow point of a wetland, and then tracer concentrations can be monitored at the outflow location. A *breakthrough curve* describes the resulting tracer concentration over time. The time required for the median concentration - when the outflow concentration equals half of the input concentration - provides an estimate of the average residence time of the system.

5.3 Wetland Water Budgets

Wetland water levels, the hydropatterns, and residence times are influenced and controlled by hydrologic inputs and outputs. In many cases, the wetland conditions observed are influenced, in large part, by the gains and losses of water. A water budget is used to account for the inputs and outputs to the wetland. The exchanges can be with the atmosphere, with ground or surface water, or by tidal action. The sum of all exchanges is what affects wetland water levels, i.e., if atmospheric exchanges cause an increase in water storage, but ground-water exchanges deplete these storages, then the total affect is the balance of the two.

The water balance equation summarizes this concept:

$$\Delta Q = I - O = \frac{\Delta V}{\Delta t} \tag{5.2}$$

where ΔQ is the difference between inflows, I, and outflows, O, and the ΔV term represents the change in water storage over a period of time, Δt. This equation means that the volume in storage in increase whenever the inflows exceed the outflows, and vice versa. Because water levels are directly related to the storage volume, and increase in storage volume always results in an increase in water levels:

$$\frac{\Delta h}{\Delta t} = \frac{\Delta Q}{A} = \frac{I - O}{A} \tag{5.3}$$

where $\Delta V = A\Delta h$, Δh is the change in water level, and A is the wetland area. This relationship holds because the change in storage equals the product of the area of the watershed and the change in water level. This relationship becomes more complicated, as noted below, whenever the water level falls below the ground surface. In these cases, the mineral and organic soil materials release less water because of their porosity and ability to retain water.

Balancing Inflows with Outflows

Potential wetland inputs include precipitation directly onto the wetland, direct overland flow, surface water inputs from rivers, streams, and marine sources, overbank flow, and from ground-water sources including subsurface, lateral unsaturated and saturated flow from uplands to toeslope and flat landscapes.

Balancing the inputs are the possible outputs, including evaporation, transpiration, ground-water recharge, and surface water outflows. Water levels rise over time when hydrologic inputs exceed hydrologic outputs, and fall when outputs exceed inputs. The change in water levels can be described using the following equation:

$$\Delta h = \frac{\Delta V}{A} = \frac{\Delta Q}{A}\, \Delta t \tag{5.4}$$

where Δh is the water level change, $\Delta V = \Delta Q\ \Delta t$ is the net change in the input water volume to the wetland, $\Delta Q = I - O$ is the net change in inputs less outputs, Δt is the time step, and A is the wetland storage area, equal to the volume of water released per unit change in water level.

Hydrologic inputs and outputs are expressed in units of volume or depth per unit time. Water levels, however, incorporate no explicit unit of time - only water level changes are expressed in terms of units of depth per unit time.

Thus, observed water levels are the result of accumulated water level changes over time and can be expressed as:

$$h(t_i) = \sum_{j=0}^{\infty} \Delta h(t_i - t_j) = \sum_{j=0}^{\infty} \frac{\Delta Q(t_i - t_j)}{A_e} \Delta t \quad (5.5)$$

where $\Delta t = t_i - t_j$. This integration of water level changes means that wetlands reduce water level fluctuations by storing water during wet periods and releasing them during dry periods. The fact that wetland water levels are accumulators of hydrologic change over time and space makes them sensitive to even small changes in environmental conditions. That is, even small alterations can manifest themselves as large changes in wetland conditions when accumulated over space and time.

In addition to a water balance equation, the mass of dissolved and suspended matter carried by the water can be balanced. The mass balance equation is written as:

$$\frac{\Delta M}{\Delta t} = \Delta L = L_1 - L_2 \quad (5.6)$$

where M is the mass of dissolved or suspended matter carried by the water, ΔM is the change in mass between two points, Δt is the time interval, ΔL is the change in *load*, L_1 is the inflow load, and L_2 is the outflow load. Clearly the rate of change in mass per unit time is a function of the balance between inflows and outflows.

Most water quality measurements are not based upon a load assessment. Instead, the solute *concentration* is normally measured. The relationship between the load, L, and the concentration, C, is found by noting that:

$$L = C \cdot Q \quad (5.7)$$

where Q is the flow rate. This is because the concentration is:

$$C = \frac{L}{Q} = \frac{M}{V} \quad (5.8)$$

or mass per unit flow rate, which is just the mass, M, per unit volume, V.

Stage-Area-Volume Relationships

Changes in water depth must normally be converted to changes in water volume. This conversion need arises because inflows and outflows are measured in terms of water volume, while water levels within the wetland are measured in terms of water depth. A conservation of mass approach can be used to equate the two quantities. The conversion from water depth, h, to water volume, V, requires knowledge of the effective storage area of the wetland, A:

$$A = \frac{\Delta V}{\Delta h} \quad (5.9)$$

where ΔV is the change volume of water and Δh is the change in water level.

For conditions when water levels are entirely above the ground surface, the water volume change per unit depth equals the wetland area. The effective storage area may change as the wetland grows in size during high stage, thus requiring the use of a table or plot of wetland stage vs. area.

Subsurface Water Storage An additional complication arises when wetland stages are below the ground surface. In this case the specific yield (i.e., the drainable porosity) of the organic and mineral sediments must be known. The specific yield is the volume of water released per unit area of wetland per unit decline in water level. In many cases, organic and mineral sediments may remain at or near saturation as water levels fall. In these cases, only a small volume of water is released from the sediments as they drain.

Combining the storage above and below the ground surface yields the following expression for the effective storage area:

$$A = A_s + S_y A_e \quad (5.10)$$

where A_s is the area of submerged wetland, A_e is the area of exposed wetland, and S_y is the specific yield of organic and mineral benthic sediments, generally equal to the difference between the saturated water content and field capacity of the sediments.

Specific yields of sediments are strongly influenced by their particle size distribution and chemical composition. Sands have large specific yields, while clays and mineral soils have low specific yields (McCuen, 1998). The specific yield can be determined by extracting core samples and determining their specific yield, or reference texts can also be consulted (Fetter, 2001; Hillel, 1971).

The resulting areas are used to generate stage-area and stage-volume relationships, from which changes in stage can be related to net changes in volume. In general, the stage-volume relationship shows a sharp change in slope once water levels fall below the ground surface, as well as when water levels overtop a natural bank or levee. Synthetic stage-area and stage-volume curves are presented in Figure 5.7.

Determining Areas and Volumes If sufficient detail is present, then wetland areas as a function of water elevation can be determined using topographic maps. Otherwise, surface mapping using a transit or level can provide cross-sections from which the volume and area can be determined. Alternatively, aerial photographs taken at different water stages can be used to estimate the stage-area relationship.

One method for determining the volume of a wetland is to add a known mass of tracer, allow the tracer to thoroughly mix in the wetland, and then measure the tracer concentration. Because the tracer concentration, C, is equal to the mass of tracer per unit volume of water,

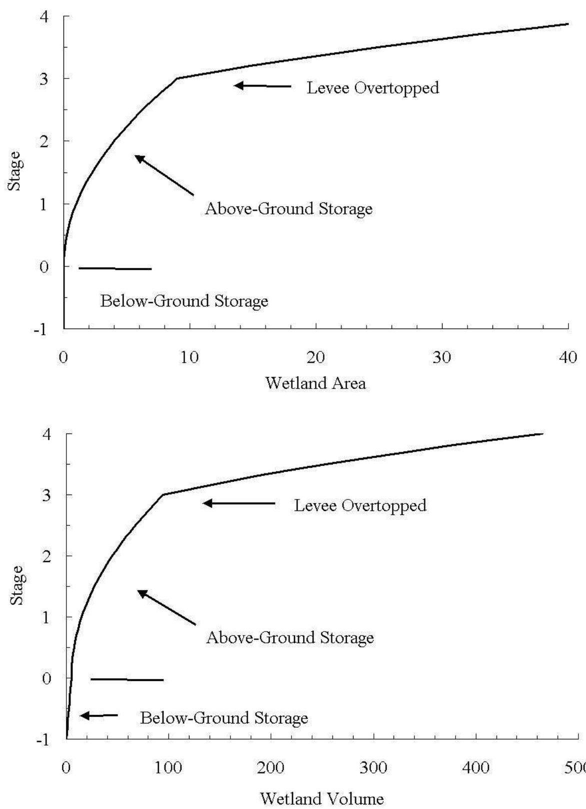

Figure 5.7: Stage-Area (top) and Stage-Volume (bottom) curves showing changes in wetland area and volume as a function of stage. Flooded wetland area is zero once water stage drops below the ground surface in the deepest section of the wetland. Volume of water storage is not zero because the bed sediments may be able to store and release water. Wetland area also increases rapidly if a levee is overtopped.

$C = M/V$, the volume of water within the wetland, V, is equal to the mass of tracer, M, divided by the tracer concentration, $V = M/C$.

If evapotranspiration is the only outflow, and there are no inflows, then the tracer concentration increases over time as the volume of water within the wetland decreases. Likewise, if precipitation or surface- and ground-water inflows are present, with no corresponding outflows, then the tracer concentration decreases over time, allowing the calculation of the wetland volume.

Water levels can be used in conjunction with the tracer data to obtain an estimate of the water balance. In effect, there are three mass balance relationships that can be used to estimate water budget components:

Water Balance $Q_1 - Q_2 = \Delta V/\Delta t$
where Q_1 is the total inflows, Q_2 is the total outflows, ΔV is the change in storage volume, and Δt is the change in time.

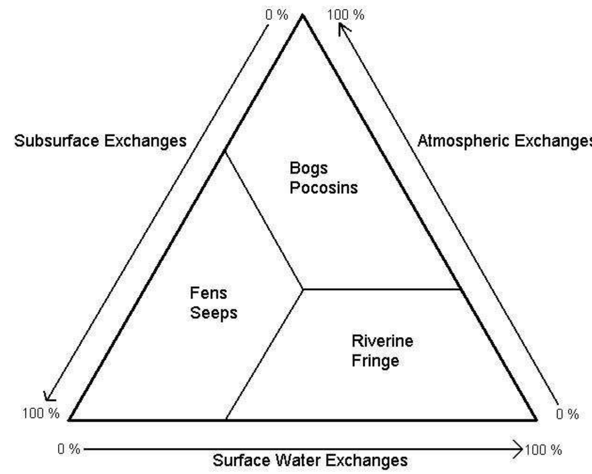

Figure 5.8: Relationship between hydrologic exchanges and nontidal wetland types.

Mass Balance $L_1 - L_2 = \Delta M/\Delta t$
where L_1 is the total mass input, L_2 is the total mass output, ΔM is the change in mass.

Concentration $C = M/V = L/Q$
where C is the concentration, M is the mass, V is the volume, L is the load, and Q is the discharge.

5.4 Water Budget Components

Water budgets are an important tool for characterizing the behavior of wetland systems. There are four general types of water sources and sinks in wetlands. The first includes atmospheric inputs and outputs, including rainfall, snow, evaporation and transpiration. The second type of water exchange includes subsurface inflows and outflows. Another exchange mechanism results from interaction with surface water, including overland flow, as well as from rivers and streams. A final type of exchange occurs in marine systems that respond to tidal variations.

Figure 5.8 presents a general characterization of wetland hydrologic exchanges for three of the four types of hydrologic exchanges (tidal wetlands are not included). The figure was adapted from Brinson (1993) which originally characterized wetlands based on the type of hydrologic inflow. The figure has been modified to show that wetlands are affected by exchanges of water - both inflows and outflows. For example, atmospheric exchanges include evapotranspiration components as well as precipitation. Surface water exchanges incorporate releases from wetlands.

As noted by Brinson (1993), wetlands have different types of inflow and outflow patterns. That is, some wetlands have simple exchanges with adjacent waterbody, such as with a river during a flood such that the wetland receives water from the river during the rising stage of the flood and returns water to the river during the falling stage. Another example are tidal exchanges along the coast, where the water moves in and out of wetlands to its original source.

In other cases, exchanges may be between different types of waterbodies and have a hybrid character. That is, inflows may be of one type (e.g., subsurface inputs) and outflows may be of another (e.g., evapotranspiration). Regardless, it is clear that identifying the key wetland hydrologic inflow and outflow components is a useful tool for understanding and managing wetlands.

Atmospheric Exchanges

Atmospheric exchanges, i.e., precipitation, evaporation and transpiration, also need to be estimated for the water budget. Rain, snow, hail, sleet, freezing rain, fog drip, dew, and frost are various forms of precipitation resulting from the condensation of tropospheric water vapor. Water levels in wetlands that are dependent on atmospheric exchanges tend to be more affected by climatic signals than wetlands dependent on ground-water sources (Orme, 1980). Lakes, like wetlands, tend to integrate climatic signals over time because of the longer residence time in these systems.

Precipitation generally occurs as discrete events, characterized by using intensity, duration, frequency, and areal extent. In aggregate, these precipitation events can be described using monthly and seasonal averages along with longer-term variability associated with climatic fluctuations. While regional precipitation networks can be used to estimate site conditions, the large spatial heterogeneity of precipitation patterns generally means that onsite precipitation measurements (either conventional or recording raingages) are needed when trying to obtain information for water balance analysis.

The loss of water from a wetland by evaporation from the water surface as well as by transpiration from plant leaf and stem surfaces can have large effects on water levels. The combined processes are called *evapotranspiration*. Evaporation dominates when open water is present, and vegetation is not. Saturated soils may lose nearly as much as open water, but not if a litter or mulch layer is present. Transpiration dominates in systems with little open water and large coverage of living vegetation. Evapotranspiration rates are affected by leaf and stem area, air, water, and plant temperatures, atmospheric humidity, wind speed, and the water potential of exposed soils.

Evapotranspiration losses from wetlands in close proximity are generally similar (Kadlec et al., 1988). This is because the source of energy for vaporization (i.e., the sun) is regionally uniform, and the availability of water for vaporization is similar, owing to the lack of water limitations in wetlands. Forested wetlands may have greater evapotranspiration rates, however, due to higher leaf areas. Also, wetlands covered with dead vegetation may have lower evapotranspiration rates due to a lack of transpiration and a reduction in evaporation from shading and poorer wind exchange. If increasing eutrophication leads to increased plant leaf area, then increased evapotranspiration water losses to the atmosphere could result.

Evaporation Theory Whether water evaporates to - or condenses from - the atmosphere is a function of the energy state of water in the liquid and gaseous forms. The vapor pressure of water (a measure of the water content of the atmosphere) is the primary measure of the energy state. Dalton's law relates the rate of evaporation to the difference in vapor pressure between the air-water interface and the vapor pressure in the atmosphere at some distance from the interface:

$$E = c\,[e_i - e_a] = c\,[RH_i\ e_s(T_i) - RH_a\ e_s(T_a)] \quad (5.11)$$

where E is the evaporation rate, c is an evaporation rate coefficient, e_i is the vapor pressure at the air-water interface, e_a is the vapor pressure in the atmosphere at some distance away from the interface, and $e_s(T_a)$ is the saturation vapor pressure, which is a function of the air temperature, T_a.

The evaporation rate coefficient, c, is a function of wind speed and the type of evaporation surface, either soil or water. The vapor pressure in the air, $e_a = RH_a\ e_s(T_a)$ equals the product of the relative humidity of the air, RH_a, and the saturation vapor pressure, e_s, based on the air temperature, T_a.

The relationship between the saturated vapor pressure and temperature for the range of liquid water at standard atmospheric pressure is:

$$e_s = 6.11\ \exp\left[\frac{17.3T}{T\ +\ 237.3}\right] \quad (5.12)$$

where e_s is the saturated vapor pressure in hPa (1 hPa = 1 millibar) and T is the temperature in $^\circ C$.

The energy state of the liquid at the air-water interface is a function of the fluid potential, or pressure, at the interface. If a free surface is present, then the fluid pressure at the air-water interface is zero, and the water potential in the atmosphere just above the surface equals the saturated partial pressure of water within the atmosphere (termed the saturated vapor pressure). In this case, the relative humidity of the air just above the interface is equal to 100% (i.e., saturated with water vapor). If the water potential at the water surface is negative, due to osmotic potentials or negative pressures within soil pores or within plant stomata, then the relative humidity above the surface is no longer saturated. The equilibrium relative humidity, RH, as a function of fluid potential, ψ, and temperature, T, is:

$$RH = \exp\left[\frac{\psi}{RT}\right] \quad (5.13)$$

where R is the water vapor gas constant.

Table 5.2: Illustrative water budget components (mm/year) for selected wetland types.

	Atmospheric		Surface Water		Ground Water	
	Precip.	ET	Inflows	Outflows	Inflows	Outflows
Southeastern Swamp	1100	1600	10	30	100	500
Northern Bog	900	700	0	200	0	0
Floodplain	1000	1400	3000	2900	300	0
Prairie Pothole	600	400	0	0	100	300

Monitoring Atmospheric Exchanges Precipitation and evaporation can be readily measured using raingages and evaporation pans, respectively. These are relatively inexpensive, and provide reliable estimates of daily atmospheric exchanges. A single raingage is usually sufficient for small wetlands (e.g., smaller than 100 ha), but multiple raingages may be required for larger wetlands, especially if significant spatial variation in rainfall is present.

Measured evaporation rates can be used to estimate evapotranspiration rates. A single evaporation pan is probably sufficient for all but the largest wetlands. While potential evapotranspiration derived from pan estimates (either manual or recording) can be used to estimate site conditions, the local effects of shading and wind shelter can adversely affect the accuracy of the measurements. Pan coefficients (the ratio of actual evapotranspiration to pan measurements) are reported to range from 0.54 to 5.3 (Carter et al., 1979).

Automated raingages are available, but more expensive than manual raingages. Automated evaporation pans are less reliable, and additional research is needed to improve their accuracy. If pan measurements are not available, then evaporation can be calculated using automated measurements of solar radiation, temperature, relative humidity, and wind speed.

Daily precipitation data should be plotted, along with daily evaporation. The difference between precipitation and evaporation can be compared to observed wetland water levels. In systems where atmospheric exchanges dominate the wetland hydrology, water levels rise during precipitation events, and fall at a rate controlled by the evaporation rate.

Subsurface Exchanges

Subsurface inflows to wetlands (also called ground-water discharge to wetlands) may result from shallow, topographically induced drainage from nearby uplands, or from discharges of regional, confined aquifers (Maley and Peters, 1999; Stuurman and de Louw, 1999). Subsurface outflows from wetlands (also called ground-water recharge) may result from downward and lateral flow from the wetland to underlying surficial aquifers, and to deeper, confined aquifers where the confining layer has been locally breached due to collapse or subsidence.

Shallow inflows may result from perched, or interflow, drainage on top of lower-permeability units within the unsaturated zone, such as clay beds, soil horizons, or even permafrost. Shallow subsurface inflows may also arise when the water surface within the wetland lies below the water table in the underlying surficial (unconfined) aquifer. In this case, the direction and magnitude of the hydraulic gradient can be estimated using aquifer water levels obtained in piezometers positioned in the vicinity of the wetland. Besides the hydraulic gradient, the water flow rate is also dependent on the permeability of the aquifer and any organic and mineral benthic sediments. The inflows may be concentrated at one or several points within the wetland. Inflows may also result from diffuse upward leakage, in which case the leakage is more uniformly distributed across the benthic materials.

Subsurface inflows from deeper sources may arise when confined aquifers discharge into the wetland. These discharges occur when the confining layer is breached due to subsidence or collapse, such as in karst areas. In this case, wetland water levels are controlled by the piezometric surface in the confined aquifer. Confined aquifer discharges can also occur when diffuse upward leakage moves through the confining layer into the overlying unconfined aquifer, and from there to the wetland. Discharges from deeper sources are less likely to respond rapidly to individual storm events, tending to respond more to seasonal and longer-term changes.

Shallow inflows may respond more rapidly to individual storm events, as well as to seasonal and climatic changes. This is because interflow and water levels in shallow aquifers tend to be more sensitive to net changes in atmospheric flux (precipitation less evapotranspiration in nearby upland areas).

Shallow subsurface outflows may occur if the wetland is underlain by a layer of low permeability that allows the water to perch. In these cases, a low point on the perimeter of the wetland allows water to exit the wetland as either overland flow, channel flow, or interflow. These systems have water levels that are perched above the regional water table, and may also have an unsaturated (vadose) zone present between the water table and the perched wetland. Recharge to the underlying, surficial aquifer also occurs through the low permeability layer.

Downward movement of water is prevented in permanently frozen soils, i.e., permafrost, because any liquid water is converted to solid form by heat exchange in the un-

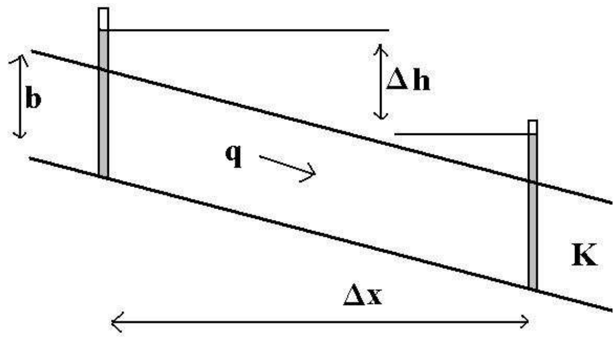

Figure 5.9: Effect of wetlands on surficial aquifer movement. Inflow to wetland is on side where water table levels are higher than wetland. Outflow is on side where water tables are lower.

derlying frozen unit. Yet, the ability of the underlying frozen unit to freeze water can be overloaded over time, resulting in a loss of the confining ability of the unit. The entire loss of the permafrost layer is possible if too much heat is added to the unit.

In some cases, wetland water levels are contiguous with water levels within the surficial aquifer (Figure 5.9). In these cases, flow through the surficial aquifer may be affected by the wetland. Normally, surficial aquifer water levels dip in the direction of water flow, while the water levels in the wetland are more horizontal. Thus, wetland water levels lie below the water table in the upgradient direction, while they lie above the water table in the downgradient direction. As a result, aquifer discharge conditions are present at the upstream end of the wetland, and aquifer recharge conditions are present at the downstream end of the wetland. This type of flow-through wetland may account for most of the flux of water through a wetland that has no readily apparent inflows or outflows.

Finally, recharge to deeper, confined aquifers may occur when subsidence or collapse has breached the confining layer that isolates the aquifer from the surface. In this case, water level increases in the wetland during wet-weather periods may cause direct recharge to the deeper aquifer (Mitsch and Gosselink, 2000).

Ground-Water Gradients The gradient of ground-water potentials governs the flow of water in the subsurface because the potential is a measure of the energy status of forces that cause water to move, and the direction of the forces controls this movement. The gradient is calculated using:

$$G = [G_x,\ G_y,\ G_z] = \left[\frac{\Delta h}{\Delta x},\ \frac{\Delta h}{\Delta y},\ \frac{\Delta h}{\Delta z}\right] \tag{5.14}$$

where G is the hydraulic gradient, composed of components in three directions, G_x, G_y, and G_z, which in turn are determined using the change in head, Δh, in each of the three directions, Δx, Δy, and Δz. Due to the layered nature of many geologic deposits, the hydraulic gradient can be simplified into horizontal and vertical components, with each layer having a unique horizontal flow pattern.

$$G_H = [G_x,\ G_y] = \left[\frac{\Delta h}{\Delta x},\ \frac{\Delta h}{\Delta y}\right] \tag{5.15}$$

and

$$G_V = [G_z] = \frac{\Delta h}{\Delta z} \tag{5.16}$$

where G_H is the horizontal component and G_V is the vertical component of the hydraulic gradient. This approach is appropriate whenever horizontal layering is present. The magnitude of the horizontal component within each layer is found by determining the change in water levels with distance, while the vertical gradient between layers is found using the change in water level with depth between two adjacent layers.

The hydraulic gradient must be combined with the hydraulic conductivity to determine the ground-water flux, or rate of volume flow. Like the hydraulic gradient, ground-water flux can be separated into three components, two horizontal and one vertical:

$$q = [q_x,\ q_y,\ q_z] \qquad q_H = [q_x,\ q_y] \qquad q_V = [q_z] \tag{5.17}$$

For flow through and across layered media, the horizontal component of flow, q_H, can be determined using the horizontal hydraulic conductivity, K_H, while the vertical component, q_V, uses the vertical conductivity, K_V:

$$q_H = -K_H\ G_H \qquad q_V = -K_V\ G_V \tag{5.18}$$

These equations are the horizontal and vertical forms of Darcy's law. The negative sign indicates that flow is from regions of higher hydraulic head to regions where the head is lower. The estimated quantities are for flow at a point. The flows must be multiplied by the cross sectional area of the unit in question to estimate flow across the area.

The total flow, Q, is calculated using:

$$Q_H = A_H\ q_H \qquad Q_V = A_V\ q_V \tag{5.19}$$

where A_H is the profile, or cross-sectional, area of flow, A_V is the map-view area of flow, and Q_H and Q_V are the horizontal and vertical components of total flow, respectively.

Monitoring Subsurface Flows Ground-water gradients - and flows - normally vary over both space and time. Thus, a high resolution of temporal and spatial sampling is required to determine the flow field with any accuracy. This means taking multiple vertical and horizontal measurements at a sufficiently frequent time interval in order to capture any variability present in the system.

Monitoring ground water flows can be accomplished by placing an array of piezometers within and around the

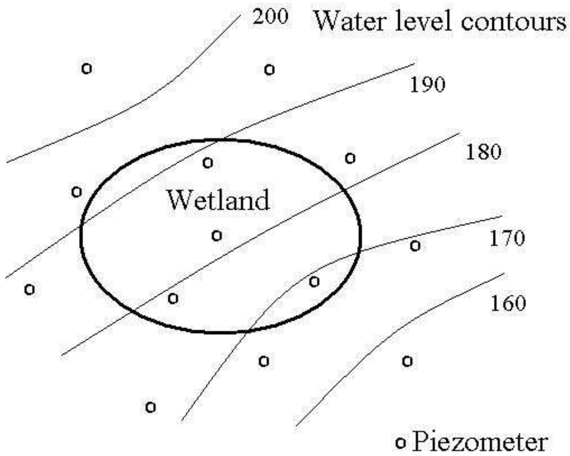

Figure 5.10: Network of piezometers required to map water levels in the vicinity of a wetland. Note that the water level contours can be made based upon interpolation between measurements within individual piezometers.

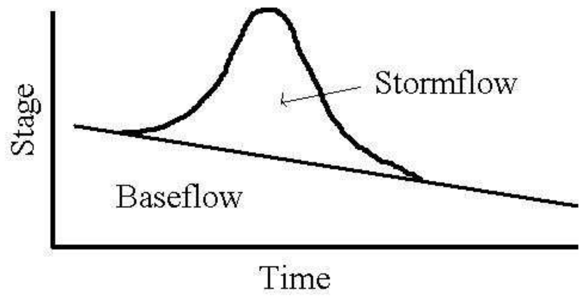

Figure 5.11: Water level behavior showing slow decline in baseflow along with a discrete storm event with its associated stormflow.

wetland. Multiple piezometers can be placed at different depths at each location to evaluate the magnitude of vertical flow. A *nest* of piezometers, composed of multiple piezometers placed at different depths, are needed whenever complex hydrostratigraphic conditions (e.g., layering) are present.

Piezometers are placed at multiple distances away from the wetland to determine the water table configuration in the neighborhood of the wetland. This network of piezometers, deployed at various locations and depths, is required to determine the three-dimensional characteristics of water potentials. Ideally, the locations of these water levels should form an equilateral triangle - not acute or obtuse triangles (Figure 5.10). Otherwise, the colinearity of the wells interferes with the estimation of the gradient.

Estimating the total flow of water into or out of the wetland requires an independent estimate of the hydrologic conductivity. Either aquifer tests can be conducted or standard tables of values can be used (Fetter, 2001). Undisturbed core samples or field testing can also be used to estimate hydraulic conductivities. Care must be taken when using core data to estimate hydraulic conductivities, in that the spatial variability of most geologic media is very high. In general, core samples tend to underestimate field hydraulic conductivities because of the difficulty in obtaining core samples for the very highest flow paths within the system.

Verry and Boulter (1979) report that the hydraulic conductivity of peat can be readily determined from its bulk density or unrubbed fiber content. They report that fibric peats have three-orders of magnitude higher horizontal hydraulic conductivity than sapric peats, due to their larger pore sizes. Daniel (1981) notes a similar effect of decomposition on hydraulic conductivity.

Once the directional hydraulic conductivities and gradients have been found, then the total flow within the system can be calculated. There are often many uncertainties with this method however, in that the spatial variability of both gradients and conductivities can be high.

Water quality sampling can be used as an independent method to determine the source of water within wetlands. For example, if shallow ground-water is moving laterally into the wetland, and ground water is also moving upward into the wetland from a deeper aquifer, then the geochemical signature of each source can be used to evaluate the relative magnitude of each inflow relative to the total.

Surface Water Exchanges

Surface waters exchanges with wetlands results from a large number of mechanisms, including overland - or sheet - flow, direct exchange when the channel of a river or stream flows through the wetland, overbank flooding during wet weather when the channel is separated from the wetland by a levee or floodplain, and along the edges of lakes, estuaries, and the ocean. These surface water exchanges result in either constant or episodic hydrologic communication between the surface water and the wetland.

Streamflow can be divided into two types, *baseflow* and *stormflow* (Figure 5.11). Baseflow is that component of flow found during low flow periods, while stormflow refers to the response to precipitation events. If a stream was flowing before the rainfall (a typical situation), stormflow is the flow that occurs in addition to the baseflow that would have occurred if it had not rained.

Flood Attenuation Not only can wetlands reduce flood velocities, they can also reduce flood wave velocities by decreasing water velocities at high discharges. Flood wave velocities travel at different rates than water velocities. The flood wave velocity, also called the flood celerity, c, is defined as the change in discharge, ΔQ, per unit change in stream cross sectional area, ΔA:

$$c = \frac{\Delta Q}{\Delta A} = \frac{\Delta(\bar{v}\ A)}{\Delta A} = \bar{v} + A\ \frac{\Delta \bar{v}}{\Delta A} \quad (5.20)$$

where $Q = A\ \bar{v}$ is the stream discharge, A is the stream cross-sectional area, and $\bar{v}$ is the mean stream velocity. This equation indicates that the flood wave velocity equals the water velocity plus a second term that is positive if the water velocity increases as the cross-sectional area (or, equivalently, stage) increases, and is negative if the water velocity slows as the area or stage increases. In other words, the wave velocity is faster than the water velocity if the water velocity increases with stage, and vice versa.

This concept can also be demonstrated using the ratio of the wave velocity to the water velocity, termed the kinematic ratio, k:

$$k = \frac{c}{v} = 1 + \frac{A}{v}\frac{\Delta v}{\Delta A} = 1 + \frac{\Delta \ln v}{\Delta \ln A} \quad (5.21)$$

This relationship again illustrates the concept that the flood wave velocity is faster than the water velocity, $k > 1$, when the second term is positive, and vice versa. The second term in the equation is the critical parameter needed to control damaging flood waves. In effect, wetlands serve to sequester flood waters, thus slowing the average and incremental water velocities, slowing flood waves travel times, and reducing peak discharges. The flood wave velocity could actually be less than the water velocity if the velocity decreases with increasing depth.

Measuring Surface Flows Surface water flows can be estimated using flow measurement control devices, such as weirs (which require a pool upstream and are not satisfactory when elevated sediment concentrations are present), flumes (which tend to flush sediments more effectively than weirs), and culverts (which are less accurate). The relationship between stage and streamflow discharge is called the *rating curve* (Figure 5.12). Once a rating curve has been developed, the stream discharge is readily found by observing the stage and then consulting the rating curve.

Open Channel Measurements Stream discharge, Q, is obtained by establishing a stream cross section, and then measuring the stream velocity, v, across the section. Because the stream velocity and depth varies across the section, the total discharge is approximated by the sum of the discharge of subsections within the total section:

$$Q = \int_A v \cdot dA \ \approx \ \sum_{i=1}^{n} v_i A_i \quad (5.22)$$

where A is the cross-sectional area of the section, n is the total number of subsections, v_i is the average velocity in each subsection, and A_i is the area of each subsection, equal to the product of the width and the average depth of the subsection.

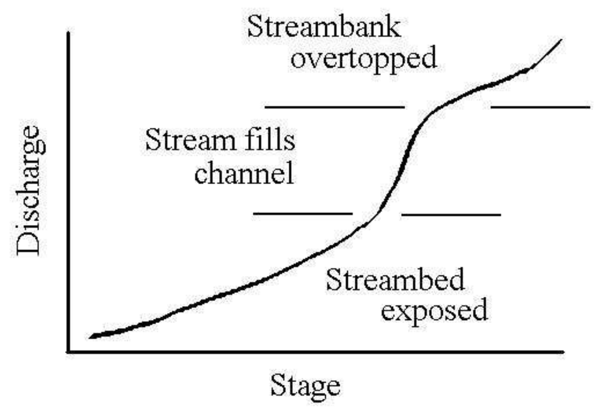

Figure 5.12: Rating curve showing relationship between water level stage and stream discharge. Note change in slope in relationship as different parts of the channel are wetted as the stage changes.

For shallow streams, the average velocity is found at approximately 60% of the depth of the stream, measured from the surface down. For deeper streams, the average velocity is found by averaging two velocity measurements, at 20% and 80% of the depth.

Upstream and downstream variations in channel shape, as well as obstructions, may cause rapid changes in velocities within the cross-section. Thus, it is important to select a location where the channel appears to be of uniform width and depth, and free of obstructions.

The site should also be selected so that *backwater* effects from downstream inflows are avoided. Another source of error occurs when the channel shape changes over time, so a solid bottom is preferred over a mobile bottom. Finally, a site located upstream of a *knickpoint* (a narrowing or shallowing of the river) is preferred over a site located downstream of a knickpoint. The knickpoint may cause *subcritical* (slow velocity) conditions upstream, and *supercritical* (high-velocity) conditions (and even a possible hydraulic jump) downstream.

To construct the rating curve, the observed stream discharge is related to the river stage, measured using a *staff gage*. The staff gage is a vertical rule placed in a protected location. Repeated measurements of discharge over a range of stages is required.

Control Structures Control structures avoid the vagaries of channel geometry by creating a uniform section. A flume can be readily constructed with a uniform cross sectional area, so that $Q = W\ h\ \bar{v}$ where W is the width of the flume, h is the depth of water in the flume, and $\bar{v}$ is the water velocity through the flume. In most cases, no unique relationship between depth and velocity can be established, being a function of the slope of the flume and the upstream and downstream conditions.

An improvement on the standard flume is to place a

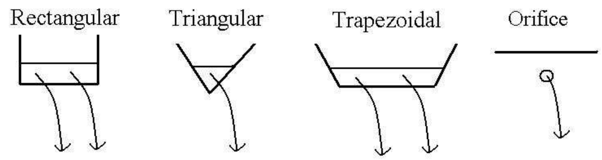

Figure 5.13: Three types of weirs (rectangular, triangular, trapezoidal, flooded orifice) used as control structures for measuring stream discharge. Water level (stage) is measured in weir basin upstream of weir blade.

Type	Weir Equation[‡]
Flooded Orifice	$Q = C\ A\ H^{0.5}$
Rectangular[§]	$Q = C_a\ W\ H^{1.5}$
Triangular	$Q = C_b\ \tan\alpha\ H^{2.5}$
Trapezoidal	$Q = C_a\ W\ H^{1.5} + C_b\ \tan\alpha\ H^{2.5}$

[†] Neglects contraction effects along weir blade edges
[‡] H is elevation of water surface in stilling basin
[§] Applies to broad-crested weirs as well

constriction in the flume (called the *throat*) that forces the flow to be subcritical (low velocity) upstream of the constriction, and supercritical (high velocity) downstream. Examples of this type include the Parshall Flume and the H-Type flume. The H-type flume was developed by the U.S. Department of Agriculture for measuring discharge in sediment-laden streams. Flumes require no upstream stilling basin, and allow sediment to pass unimpaired through the structure. Ice, leaves and other debris can still affect the reading, however.

Yet another control structure is the weir (Figure 5.13). A weir has a stilling basin upstream of a constriction (normally called the *weir blade*), and a free-fall below the constriction. The stilling basin is used to eliminate the velocity head, yielding $H = z$ in the stilling basin, where z is the measured water surface elevation above the lowermost point on the weir blade.

Two general types of weirs include *broad-crested* and *sharp-crested*. As the name implies, the broad-crested weir has a broad constriction in the direction of flow, while the sharp-crested weir has a knife-edge blade that forms the constriction. A broad-crested weir consists of an outflow structure over which water flows for some distance before falling over the downstream edge. A sharp-crested weir is constructed so that the flowing water passes over a vertical, knife-edge, thus minimizing resistance with the weir blade.

Water flows out of the weir over the weir blade, which can take a variety of shapes, including triangular, rectangular, and trapezoidal. A submerged, circular orifice can also be used. The discharge is calculated using $Q = vA = v\ (WH)$ where $A = WH$ is the cross-sectional area perpendicular to flow above the weir blade, W is the width of the weir, and H is the depth of flow above the weir blade. So that the general weir formula is:

$$Q = a\ H^b \tag{5.23}$$

where a accounts for the cross-sectional area as well as contraction and energy losses, H is the water surface elevation in the stilling basin above the weir blade, and $b = 0.5$ for a flooded orifice, $b = 1.5$ for a rectangular weir, and $b = 2.5$ for a triangular weir. This equation holds for the broad-crested weir as well, with $b = 1.5$.

Weirs may not provide accurate estimates in several situations. One source of error occurs when the weir blade becomes blocked by ice or floating debris, such as leaves and branches. Another source of error arises when the weir basin fills with sediments, resulting in an inaccurate estimate of the total head. For all weirs, a staff gage or a water-level recorder is placed upstream of the constriction to measure the total head under subcritical conditions. Weirs tend to be more accurate than flumes, but suffer from sediment accumulation in the stilling basin, along with debris obstructing on weir crest.

Culverts under roads can also be used as control structure. Four types of flow conditions can be found for most culverts; upstream intake either submerged or open, and downstream discharge conditions either submerged or open. When the upstream end is flooded, and the downstream end is open, then the orifice solution for the weir equation can be used

When both upstream and downstream opening are submerged, then pipe flow conditions are present, and discharge can be found using the difference in head between the two ends, the culvert length and diameter, and the type of culvert (smooth, corrugated, etc.). When the upstream end is open and the downstream end is either open or flooded, then the flow can be found using indirect techniques such as the Manning's equation, below.

Regardless of flow conditions, it is better if the culvert has a uniform shape throughout its length and is not obstructed with debris. Elevation measurements can be made mechanically using a water level recorder, or visually using a staff gage.

Indirect Measurements For situations when control structures are not present, and instream measurements are not possible, then Manning's equation is commonly employed to indirectly measure water velocity:

$$\bar{v} = \frac{1}{n}\ R^{2/3}\ S^{1/2} \tag{5.24}$$

where n is the Manning's roughness coefficient, R is the hydraulic radius, and S is the gradient in total head. The roughness coefficient is normally found in tables, and is based on stream channel characteristics such as stream bed materials, amount of vegetation within the channel, variation in channel shape, and sinuosity. The hydraulic

Table 5.3: Types of Channel Control Structures

Weirs:

- stilling basin is located upstream of weir
- water level recorder is used to measure stage in stilling basin
- outlet structures include rectangular, triangular (v-notch), and Cipolletti (trapezoidal) shapes
- weir crests can be broad (flat lip) and sharp (knife-blade) crested
- flow is subcritical upstream of crest, supercritical downstream
- weirs collect sediments in the stilling basin, debris on weir blade

Flumes:

- no stilling basin, only a narrow throat
- regular approach section
- passes sediment easily
- woody debris can be a problem

Culverts:

- Four combinations of flow equations, flooded vs. open upstream, flooded vs. open downstream
- Culvert should be a regular shape, round or rectangular, with no debris

radius is defined as the ratio of the stream cross-sectional area to the wetted perimeter, $R = A/P$, which is approximately equal to the water depth in a shallow, wide channel. The total head gradient is the drop in total head per unit distance of stream channel.

Once the average water velocity, v, has been determined, the stream discharge, Q, can be estimated using:

$$Q = \bar{v}A \tag{5.25}$$

where A is the cross-sectional area of the channel.

Estimating Overbank Flows Wetlands adjacent to riverine systems are often affected by overbank flows during stormflow periods. In these cases, the river spills out of its normal channel and overflows onto adjacent floodplains. The period of time that wetlands on the floodplains are affected by the duration of stormflows, and their amplitude. Instrumentation to monitor water levels in wetlands adjacent to riverine systems can be installed using techniques mentioned previously. Additionally, the U.S. Geological Survey provides estimates of overbank flooding frequencies for ungaged sites (Jennings, et al., 1994).

Tidal Exchanges

Coastal wetlands are similar in many ways to freshwater wetlands, except that they are transitional between marine and terrestrial environments. Coastal wetlands have unique attributes that distinguish them from both terrestrial and marine systems. It is, in fact, the combination of flooding and soils near the water surface that promotes ecologic diversity and productivity in coastal wetlands.

Coastal wetlands occupy similar landscape positions as lacustrine wetlands. Great variability exists, however, among coastal wetlands. Some coastal wetlands are dominated by the ebb and flow of water levels of the ocean due to tides, termed *tidal* wetlands. The large magnitude of the daily tides, along with their regularity, result in unique wetland conditions. Other marine wetlands are more sheltered from tidal effects. Still others may be affected by water quality changes resulting from freshwater tributaries.

Tidal Effects Many coastal wetlands function to dampen tidal and wave energies. Because viscous and turbulent drag through coastal wetlands dissipate energy, the magnitude of fluctuations generally diminishes with distance from the coast. The equation for a harmonic wave height with constant energy dissipation as a function of distance is (Carslaw and Jaeger, 1959, Eq. 2.6.8):

$$h = h_o \; e^{-\kappa x} \; \cos(\omega t - \kappa x) \tag{5.26}$$

where $\kappa = \sqrt{\omega/2D}$, h is the wave or tidal height, h_o is the maximum magnitude of the fluctuation at the shoreline, κ is a damping coefficient that accounts for the dissipation of energy with distance, x is distance from the shoreline,

ω is the frequency of the fluctuation, t is time, and D is the hydraulic diffusion coefficient. The amplitude of the oscillation, $|h| = h_o\ e^{-\kappa x}$, decreases with increasing frequency and distance from the shoreline, as does the lag, κx. This means that low-frequency waves, such as daily and twice-daily tidal fluctuations, propagate deeper into coastal regions, are attenuated less and lagged more, than high-frequency waves.

While the harmonic wave height equation may not be suitable for many coastal areas, it does combine most of the important factors (e.g, time, distance, wave frequency, and hydraulic resistance) that affect water and energy movement. An understanding of the local coastal morphology and vegetation is important, along with the energy inputs from marine sources, as well as from terrestrial surface-water and ground-water inflows (Dronkers, 1986). Together, these external factors influence how coastal wetlands are affected by hydrologic tidal exchanges.

Tides can affect wetlands many miles from any sources of saline water. The hydrodynamic conditions imposed by a changing sea level cause rivers to flow more slowly during high tides, and vice versa. Thus, water levels in rivers upstream of the coast rise and fall in tandem with the tides. The magnitude of this effect is diminished with distance upstream, and is a function of local channel features.

Other Inputs Many coastal wetlands are affected by nearby freshwater inputs, especially in estuarine environments. In these cases, occasional, large stormwater inflows can cause rapid changes in the salinity, temperature, dissolved oxygen, and sediment concentration within the wetland. In some cases, these inputs can be beneficial, such as historical sediment deposition in the Mississippi River delta region of southern Louisiana - in contrast to the current practice of diverting stormflows away from coastal wetlands which has led to regional subsidence and salt water intrusion. In other cases, these inputs can be detrimental, as when increased urban wastewater and stormwater inputs to coastal estuaries alters the natural conditions.

Ground-water inputs to coastal wetlands can also be significant, and can take two forms, point and diffuse. *Point* discharges - such as springs - form when an underlying confining layer for an artesian aquifer is breached, allowing the upward flow of water. *Diffuse* upward leakage occurs when the confined artesian aquifer discharges over a large region. In these cases, the leakage is more spatially uniform, with greater amounts of leakage occurring in low-elevation areas. Ground-water exchanges may be more difficult to characterize than other sources, however. One promising method is the use of radioisotopes to differentiate between groundwater and other inputs (Charette et al., 2003).

Monitoring Coastal Wetlands In some cases, distinguishing tidal from freshwater and ground-water inputs can be achieved using geochemical information. The tidal water quality is clearly distinguished by their high concentrations of sodium-chloride type water, while freshwater inputs normally have markedly lower specific conductivities and total dissolved solids. Depending upon location, ground-water inputs are intermediate, with possibly distinct geochemical signatures. For example, if carbonate aquifers are present, then a calcium signal may be present in these waters.

The degree of tidal flushing can thus be monitored using water quality data to characterize the residence times of water within the system. Estimates of fluxes can then be estimated based on the water balance equation.

5.5 Evolution and Alteration of Wetland Hydrology

Wetlands change over time. Natural processes - such as sediments that fill wetlands and beaver activity - as well as accelerated processes - such as upstream development and direct alteration of the wetland - all cause changes that affect wetland hydrologic behavior. This section discusses some of these effects, focusing on some of the many factors that cause wetlands to change over time.

Natural Forces of Change

Wetlands are formed as the result of many geologic forces. Rivers form flood plains that provide a landscape position that enhances wetland development. Glaciers scour the landscape, leaving behind features that also promote wetlands. Tectonic uplift and subsidence create depressional features that are favorable to wetland formation. Carbonate aquifers dissolve over time, leaving behind depressions where wetlands can form. Also, accelerated erosion transports sediment out of natural channels, leading to down-cutting and deepening of channels, which leads to a lowering of riparian water tables and the reduction of overland flows, both of which alter wetland saturation.

Wetlands can modify their environment as they mature. Peats may substantially modify the original landscape by filling in the depression they originally formed in (Daniel, 1981). Other biological forces also promote wetland formation. Beaver create impoundments which form natural wetlands in habitats that are favorable to their needs. Large woody debris also form natural dams that impound shallow wetlands.

Once formed, wetlands can also age over time, slowly filling in with external sources of materials, such as sediments from upland erosion, as well as with detrital materials from wetland vegetation. Rates of deposition of these materials can be slow, such as in oligotrophic systems with small upstream catchment areas. Or they can be rapid, such as in nutrient-rich areas with large upstream areas with extensive erosion.

Wetlands in a natural setting, therefore, are constantly being formed and lost - depending on the balance of forces.

Wetland hydrology therefore also changes over time. Reducing the volume of storage within a wetland decreases the residence time, and can also reduce the depth and hydropattern by removing storage volume within the deepwater areas that would normally remain wet under drought conditions.

The dynamic nature of hydrology - especially when applied to wetlands - means that wetlands can not be investigated apart from their regional environment. Hydrologic alteration upstream of the wetland affects wetland evolution.

Human Alteration of Wetland Hydrology

Humans have substantially increased hydrologic disturbances within watersheds (Azous and Horner, 2001; Fisk, 1989). These changes generally cause increased sediment production and transport, as well as increases in nutrient concentrations and loads. Such increases naturally cause reductions in wetland storage volumes due to sediment trapping and nutrient uptake with subsequent deposition of organic sediments.

Surface water inflows to wetlands can be increased by routing stormwater runoff into them from urban, industrial, and agricultural areas. Inflows are also altered when hydraulic structures, such as reservoirs, canals, levees, dikes, revetments, and jetties obstruct or alter natural hydrologic patterns. Many of these alterations resulted from efforts to drain wetlands.

Outflows from wetlands were increased by the construction of drainage ditches, channels and canals, or the removal of natural barriers, such as vegetation, and by straightening streams. Other efforts to drain wetlands used ground-water extraction techniques, such as underground tile drains and pumping wells that lead to lowered ground-water levels. Lowering of water tables can affect wetlands by increasing subsurface drainage from the wetland to the point of ground-water extraction. Ditches and tile drains increase the discharge of shallow ground water, thus lowering water tables in the vicinity of the drain. Water levels increase with distance away from the drains, reaching a maximum midway between the drains. Tile drainage systems increase the rate of shallow ground-water flow, thus favoring drier conditions within the wetland. Tile drainage systems are more effective for removing water resulting from low-intensity, long-duration storms, and are less effective for draining water resulting from short-duration, high-intensity storms (Galatowitsch and van der Valk, 1994).

Drainage of fields for agriculture may reduce surface water inflows, lower water tables, an reduce the seasonal period of soil saturation. Ground-water pumping in the vicinity of the wetland can lead to a reduction in shallow aquifer water levels, while irrigation may increase water levels, resulting in either decreases or increases in wetland water levels, respectively. The effects of regional ground-water pumping tend to manifest themselves as slow (and in some cases, rapid) declines in regional, confined aquifer levels. These declines are then transmitted by reductions in diffuse upward leakage or direct connections to wetlands, resulting in the lowering of water levels in wetlands.

The effect of the alteration of channels and canals can be two-fold - not only do the new excavations convey more water out of the wetland, the *spoils* (i.e., the materials removed from the excavated areas) are commonly piled near the excavations and may concentrate or otherwise alter the natural drainage through the wetland (Chabreck, 1988).

As mentioned above, beaver ponds form natural wetlands that once dotted the landscape. Beaver trapping and eradication efforts may therefore have reduced the formation of new wetlands. Also, reducing the availability of large woody debris may also reduce wetland formation. Harvesting of riparian vegetation - particularly the larger diameter trees - could result in poorer recruitment of large woody debris. While wetlands clearly affect vegetation, fish, and wildlife, it is also true that these biological factors affect wetlands.

Obstructions (such as beaver dams, roads, channels, dams) to surface water exchanges alter the hydrology by requiring a higher stage to pass the same flow. Obstructions to inflows may deprive the wetland of natural flows. In some cases, obstructions may not substantially alter total wetland outflows, they may just alter the stage-discharge relationship, requiring a higher water level in order to pass an equivalent discharge. This may have both positive and negative effects on wetlands. Increased water levels can alter the natural storage ability (a negative effect), but may increase the residence time (a positive effect).

Some wetlands can have a large hydraulic effect by mitigating flood flows. Wetlands can retard floods by slowing the average water velocity as well as the flood peak velocity. Rapid flood waves cause greater damage downstream because they have less time to dissipate their peaks. Slower floods have smaller peaks and lower velocities, resulting in less downstream damage. This is accomplished by the additional friction, or resistance to flow, that wetland provide, along with the increased water storage capacity associated with their area.

Efforts to mitigate stormwater runoff have resulted in efforts to design and construct artificial wetlands to mimic the beneficial hydrologic effects of natural wetlands (Kadlec et al., 1993; Walker, 1995; Walker and Kadlec, 2002; Moustafa and Kadlec, 2002). These efforts focus on using hydrodynamic models to achieve specific management goals that require reductions in nutrients, sediments, and peak flows.

As noted earlier, Manning's equation suggests that several factors affect the water velocity, including the flow roughness of flooded ground, the hydraulic radius (i.e., the effective water depth), and the water energy slope. Wetlands with substantial macrophytic vegetation can increase the hydraulic roughness, thus decreasing flow velocities. Also, shallow water bodies reduce the hydraulic radius, again decreasing flow velocities. Removing wetland vegetation thus decreases the hydraulic radius, causing increased water velocities.

It is also apparent that increased wetland loading rates, along with decreased retention times, substantially decreases the effectiveness of wetlands in storing water during flood periods and subsequently releasing them during dry periods. The resulting hydrologic performance of affected wetlands is fundamentally compromised.

Irrigation can increase water tables if the amount of irrigation exceeds plant needs. In many arid areas, surplus irrigation is required to remove dissolved salts from the root zone. Under these conditions, recharge from irrigation, precipitation, and/or surface flows may increase water levels to the point where surface inundation results, forming saline wetlands (Mitsch and Gosselink, 2000).

As noted earlier, some coastal wetlands are dominated by nearby freshwater inputs, especially in estuarine environments. In these cases, occasional, large stormwater-driven inflows can cause rapid changes in the salinity, temperature, dissolved oxygen, and sediment concentration within the wetland. Alteration of the coastal morphology by dredging can adversely affect natural wetlands by increasing saltwater intrusion rates (Chabreck, 1988). Density-dependent stratification of estuarine waters may prevent salt-water presence in coastal areas. Construction of deep-water navigation channels may allow for salt-water to gain inland access, which can result in increased salinities.

Coastal ground-water pumping affects coastal wetlands by reducing artesian pressures in underlying confined aquifers, which may then cause a reduction in point and diffuse upward leakage. Also, ground-water pumping may cause coastal subsidence, resulting in the effective lowering of the ground surface relative to the sea level, causing the intrusion of saline water into coastal wetlands.

Pumping from shallow aquifers can also lower coastal zone water levels, causing local dewatering of coastal wetlands. Shallow disposal of septic wastes can alter local ground-water quality by increasing organic and nutrient loading, and decreasing dissolved oxygen concentrations. These changes can affect local wetlands when and if this ground water discharges into them.

Wetland Restoration

Efforts toward restoration of the hydrologic function of compromised wetlands are currently expanding (Hey and Philippi, 1999; Means and Hinchee, 1999). Additional efforts are being undertaken to create artificial wetlands that take advantage of natural functions that wetlands provide (Kadlec and Knight, 1995; Hammer, 1996). Regardless of whether impaired wetlands are being restored, or new wetlands are being created, the intent is to recreate the hydrologic behavior that we find so important (Greeson et al., 1979). The emphasis in these cases is the design and evaluation of alternative strategies for wetland restoration (Marble, 1992).

As shown earlier, water levels in wetlands can be controlled by manipulating the stage-discharge relationship. Changing the elevation of an outflow structure, e.g., by raising the base of an outlet weir elevation, changes the wetland water levels and flooded areas. The base elevation along with the rate of change in discharge with elevation can be adjusted using outflow structures of different sizes and shapes, depending upon the desired outflow characteristics.

Other hydrologic alteration possibilities include closing of ditches and drains, thus reducing wetland outflows. Removing artificial obstructions, such as roads and berms can also improve flow through the wetland by recreating natural hydrologic communication with neighboring waterbodies.

These principles apply not just to freshwater wetlands, but also apply to the restoration of tidal wetlands, which requires the recreation or simulation of natural hydraulic conditions (Zedler, 2001). Weirs and plugs are devices used in tidal marshlands to maintain minimum water levels (Chabreck, 1988). Weirs are useful because the bottom elevation of the weir controls the minimum elevation on the upstream side, but allows higher flows to pass unaffected. Ditch plugs provide the same control, but are more susceptible to destruction during high flows.

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