INTRODUCTION
Lakes and ponds occur in a wide range of depths, sizes, and permanence—from deep lakes having a permanent body of surface water to shallow ponds having water for only a few weeks each year. These factors also vary within each lake or pond, resulting in diverse communities of aquatic plants growing in various patterns. Certain types of plants require relatively...
high water levels, while others cannot tolerate standing water. Therefore, water level change is considered a disturbance to many aquatic plants.

Dynamic changes in water level are controlled by the balance between inputs and outputs of water, which are in turn controlled by the hydrological processes. Many hydrological processes are sensitive to changes in climate. For example, during a prolonged drought, precipitation inputs generally decrease and evaporation outputs increase, resulting in a drawdown of lake level or even a complete drying out. Climate also affects the lake water balance by changing the amount of stream flow and groundwater flow into the lake, but the response of the hydrological processes to climate is complicated because of the complex interactions among climate, vegetation, soil, and groundwater. Such interactions are also strongly affected by land-cover change caused by natural (e.g., fire) or anthropogenic (e.g., agriculture) processes.

This chapter discusses the hydrological processes that control the water level change of surface water and shallow groundwater. The objective is to present essential hydrological principles and practices so that ecologists and hydrologists can engage in meaningful interdisciplinary research. We first present the concept of the water balance, followed by a discussion of individual components of the water balance. We also discuss the effects of climatic fluctuation and landuse change on the water balance by using examples from prairie wetland ecosystems, with particular emphasis on the ecohydrological linkage between water and riparian vegetation.

WATER BALANCE

Water Balance Equation

The water level in a lake is controlled by the balance between input and output:

\[ Q_{in} - Q_{out} = A \frac{dh}{dt} \]  

(1)

where \( Q_{in} \) [L\(^3\) T\(^{-1}\)] is the sum of all water inputs, \( Q_{out} \) [L\(^3\) T\(^{-1}\)] is the sum of all outputs, \( A \) [L\(^2\)] is the surface area of the lake, and \( \frac{dh}{dt} \) [L T\(^{-1}\)] is the rate of water-level (h) change. Note that A represents the water-covered area, which is normally dependent on h (the higher the water level, the greater the surface area). The water regime of a lake (i.e., how A and h change over time) is determined by the seasonal and interannual variability of \( Q_{in} - Q_{out} \). Therefore, understanding the water regime requires some knowledge...
of the hydrological processes controlling $Q_{in}$ and $Q_{out}$. The inputs include perennial and intermittent streams, groundwater inflow, direct precipitation onto the lake, diffuse runoff from the shoreline, and snowdrift. The outputs include streams, groundwater outflow, and evaporation and transpiration, commonly called evapotranspiration (Fig. 1). In addition to these natural processes, $Q_{in}$ and $Q_{out}$ may include artificial terms, such as water intake for irrigation or wastewater discharge.

**Precipitation**

Direct precipitation is an important component of the water balance. Depending on the atmospheric conditions, precipitation occurs as rain, snow, hail, or various other forms. Rainfall is relatively easily measured in principle and, hence, is usually the most accurately measured term in the water balance equation. However, precipitation may have significant spatial variability, even over a relatively small area. Therefore, accurate determination of event-by-event precipitation inputs requires multiple precipitation gauges distributed over the study area. Winter precipitation in cold regions mostly occurs as snow. Snowflakes are easily moved by wind, making it difficult to measure snowfall accurately, even with gauges equipped with windshields (Goodison et al., 1998). Therefore, one should be aware of the uncertainty associated with measured winter precipitation, even though such uncertainty may be much smaller than the uncertainty regarding other terms in the water balance. Many hydrological studies only require periodic measurements of the amount of water equivalent in the snowpack. In these studies, the average snow water equivalent over a relatively large area is estimated from measurements of snow depth and density along representative survey lines.

Part of the precipitation is intercepted by leaves and stems of emergent plants and returned to the atmosphere by evaporation. This process can be important for the water balance of shallow marshes having a sizable area
covered by plants. The amount of interception is controlled by the leaf area index, the total area of all plant leaves per unit area. Leaf area index is commonly estimated from the comparison of radiation measurements beneath and above the canopy because direct determination of the index is impractically labor intensive (Weiss et al., 2004). Interception is also controlled by the shape and angle of leaves. Flat leaves oriented horizontally can intercept precipitation more effectively than narrow leaves oriented at steep angles (Crockford and Richardson, 2000).

Emergent vegetation also affects snow accumulation. Once deposited on the ground or frozen lake surface, the snowpack typically remains frozen for several days to months, allowing wind-driven snowdrifts to redistribute snow across the landscape (Pomeroy et al., 1998). Tall grasses or shrubs effectively trap drifting snow and serve as a snow “sink.” Where vegetation is nearly absent, such as on cultivated fields, snow tends to accumulate in topographic depressions and, upon melting, releases a major water input to upland ponds and wetlands that form in these depressions (van der Kamp et al., 2003).

**Evapotranspiration**

Water is transferred from ponds and lakes to the atmosphere by direct evaporation from the water surface and transpiration by emergent plants. The two processes are driven by the same meteorological factors and are commonly lumped together as evapotranspiration (ET). The phase change of water from liquid to vapor is an essential part of ET. Therefore, ET can be understood in context of the energy exchange at the lake surface. However, the phase change needs to be accompanied by removal of vapor from the lake or leaf surface by the turbulent mixing of air. Therefore, ET can also be considered an aerodynamic mass-transfer process. Not surprisingly, most of the equations that best model ET over lakes and wetlands contain both energy and aerodynamic terms (Winter et al., 1994; Rosenberry et al., 2004). A comprehensive review of methods to estimate wetland ET is found in Drexler et al. (2004).

The primary source of energy for ET is radiation inputs. The wavelength of radiation is dependent on the temperature of the source. Solar radiation has much shorter wavelengths than the radiation emitted by the atmosphere and by the earth’s surface. Therefore, solar radiation is commonly referred to as shortwave radiation, and the radiation emitted by lower-temperature surfaces is referred to as longwave radiation. The lake surface receives solar radiation and reflects part of it back. The lake surface
also receives and emits longwave radiation. Water is an excellent absorber of shortwave radiation and allows only a small portion of incoming shortwave radiation to be reflected back. The ratio of reflected to incoming shortwave radiation (albedo) is typically about 0.03 to 0.05 for water surfaces. The albedo becomes much higher when the lake surface is partially covered by emergent plants or ice. The amount of longwave radiation emitted by a water surface \((Q_{LW} \uparrow [M \cdot T^{-3}])\) is proportional to the fourth power of the temperature of the surface \((T_s[K]):\)

\[
Q_{LW} \uparrow = \varepsilon \sigma T_s^4
\]

where \(\varepsilon\) is the emissivity of the water surface (typically 0.95 to 0.98) and \(\sigma\) is the Stefan-Boltzmann constant \((5.7 \times 10^{-8} \text{ J m}^{-2} \text{s}^{-1} \text{K}^{-4})\). Therefore, outgoing longwave radiation is strongly dependent on the lake surface temperature. The net radiation \((Q_n [M \cdot T^{-3}])\) received by the surface is obtained by adding all incoming radiation and subtracting all outgoing radiation. Part of \(Q_n\) is stored in the water in an amount proportional to the total water volume times the change in average water temperature, while the rest is used to convert liquid water molecules to vapor and to warm the air in contact with the water.

The vapor-holding capacity of air increases with temperature (Lowe, 1977). A parcel of air saturated with water vapor has a relative humidity of 100%. A thin layer (<1 mm) of air immediately above the water is saturated with water vapor, and its temperature is equal to the water-surface temperature \((T_s)\). Therefore, the partial pressure of water vapor \((e_s\) in this layer is uniquely determined by \(T_s\). The rate of increase in saturation vapor pressure with respect to temperature has a special significance in evaporation and is represented by \(\Delta [\text{mb K}^{-1}]\). Suppose that the air temperature \((T_a\) at some distance, say 0.1 m, above the lake surface is lower than \(T_s\), then mixing the warmer air, which is in contact with water, with the colder air above will result in the net upward transfer of energy. This process and the reverse, transfer of energy from warm air to cold water, is called sensible heat transfer, with a positive quantity \((Q_h [M \cdot T^{-3}])\) assigned for upward flux. Suppose now that the air at 0.1 m is relatively dry so that the vapor pressure \((e'_a\) of this air is less than \(e_s\); mixing more humid air, which is in contact with the water, with the less humid air above results in the net upward transfer of vapor. This process is called latent heat transfer because water vapor carries with it the latent heat of vaporization. A positive quantity \((Q_e [M \cdot T^{-3}])\) is assigned for upward flux so that the evaporation rate \((E [L \cdot T^{-1}])\) is directly proportional to \(Q_e\):
\[ E = \frac{Q_e}{(\rho_w L_e)} \]  

(3)

where \( L_e \) is the latent heat of vaporization (\( 2.47 \times 10^6 \) J kg\(^{-1}\) at 293 K) and \( \rho_w \) is the density of water (1000 kg m\(^{-3}\)). A similar argument applies to the sensible and latent heat transfers between a leaf surface and ambient air. Since turbulent mixing is driven by the motion of air, it is reasonable to expect that \( Q_h \) and \( Q_e \) increase with wind speed. It is also reasonable to expect that \( Q_h \) and \( Q_e \) are dependent on the gradients of temperature and vapor pressure. Numerous meteorological studies have established empirical relations between mean horizontal wind speed (\( u \) [L T\(^{-1}\)]) and \( Q_h \), and between \( u \) and \( Q_e \):

\[ Q_h = \rho_a C_p f(u)(T_s - T_a) \]  

(4)

\[ Q_e = \rho_a \left( \frac{0.622}{p_b} \right) L_e f(u)(e_s - e_a) \]  

(5)

where \( \rho_a \) [M L\(^{-3}\)] is the density of air, \( C_p \) [L\(^2\) T\(^{-2}\) K\(^{-1}\)] is the specific heat of air, and \( p_b \) [M L\(^{-1}\) T\(^{-2}\)] is barometric pressure. The constant 0.622 is the mass ratio between the molecules of dry air and water vapor (Arya, 1988). The wind function \( f(u) \) has a dimension of [L T\(^{-1}\)] and represents the empirical relation between \( u \) and the degree of turbulent mixing. The most commonly used form of wind function is linear, such as:

\[ f(u) = a + bu \]  

(6)

where \( a \) [L T\(^{-1}\)] and \( b \) are empirical constants. The same wind function is commonly used for both sensible and latent heat transfers, assuming that the two processes are driven by the same mixing mechanism. If \( a \) and \( b \) are known, ET flux can be estimated from Equations (3) and (5) by using the observed \( u \), \( e_s \), and \( e_a \). This is commonly called mass-transfer or Dalton-type formula in the literature (Bras, 1990).

In a well-developed boundary layer over a homogeneous surface, \( u \) is essentially zero at some short distance (\( z_0 \)) above the land surface and increases with the logarithm of height (Arya, 1988). This distance is called the roughness length and has a strong influence on \( a \) and \( b \) in Equation (6). As the name suggests, \( z_0 \) is dependent on the physical roughness of the surface, such as that caused by waves or emergent plants. Dense growth of tall plants such as Typha or Phragmites can result in \( z_0 \) being substantially above the water surface. In addition to \( z_0 \), thermal stratification of the air can also have a major influence on \( a \) and \( b \), especially when \( u \) is relatively
small. The atmosphere above a lake surface is said to be stable when a cold layer of air is overlain by warm air because the upward motion of colder, denser air into warmer, lighter air is countered by gravity. In contrast, the atmosphere is unstable when a warm layer of air is overlain by cold air. Since $a$ and $b$ are dependent on so many variables, they need to be determined empirically for each site. Therefore, mass-transfer formulas are usually used in combination with energy balance equations described below.

The energy balance of a lake is given by:

$$Q_n - Q_g + Q_a - Q_w = Q_b + Q_e$$

where $Q_n$ is net radiation averaged over the lake area, $Q_a [M \cdot T^{-3}]$ is the net energy input by stream and groundwater divided by the lake area, $Q_w [M \cdot T^{-3}]$ is the area-averaged rate of energy storage change in lake water, and $Q_g [M \cdot T^{-3}]$ is the area-averaged energy flux into the lake bottom sediment. $Q_a$ is usually a minor component for most lakes and ponds, but it may be important for systems with artificial inputs of warm water. In principle, $Q_w$ is given by:

$$Q_w = \frac{\rho_w C_w}{A} \frac{dT}{dt}(VT_w)$$

where $\rho_w$ is the density and $C_w [L^2 \cdot T^{-2} \cdot K^{-1}]$ is the specific heat of water, $A$ is the lake area, $V$ is the total volume of water, and $T_w$ is average water temperature. Determining $T_w$ may be difficult because it requires monitoring the depth-temperature profile at multiple locations (Stannard and Rosenberry, 1991). However, in some lakes, measuring the depth-temperature profile at only one location will result in very little loss of accuracy (Rosenberry et al., 1993). The sediment heat flux ($Q_g$) may be estimated from temperature gradients and the thermal conductivity of sediments. Its magnitude is generally smaller than the magnitude of $Q_w$, except for very shallow ponds and wetlands.

The left-hand side of Equation (7) represents the available energy for turbulent heat transfer and can be lumped together into a single term $Q_{b+e}$:

$$Q_{b+e} = Q_n - Q_g + Q_a - Q_w$$

For large lakes with negligible “shoreline effects” (described below), $Q_e$ can be calculated from $Q_{b+e}$ and the Bowen ratio ($\beta$) given by:
\[ \beta = \frac{Q_h}{Q_e} \]  

(10)

It follows from Equations (4) and (5) that:

\[ \beta = \gamma \frac{(T_s - T_a)}{(e_s - e_a)} \]  

(11)

where \( \gamma \) is the psychrometric constant defined as

\[ \gamma = \frac{C_p \ p_b}{(0.622 \ L_e)} \]  

(12)

and takes a value of about 0.66 mb K\(^{-1}\) at \( T_a = 293 \) K and \( p_b = 1000 \) mb.

From Equations (7), (9), and (10), \( Q_e \) is given by

\[ Q_e = Q_{h+\varepsilon} / (1 + \beta) \]  

(13a)

The Bowen ratio energy balance method uses Equations (11) to (13a) to estimate ET flux (e.g., Burba et al., 1999). Equation (13a) can be written in a different form called the Priestley and Taylor (1972) equation:

\[ Q_e = Q_{h+\varepsilon} \ \alpha \Delta / (\Delta + \gamma) \]  

(13b)

where \( \alpha \) is a dimensionless coefficient and \( \Delta \) is the slope of the saturation vapor pressure-temperature curve. When data are averaged over a period longer than several days, many studies have found that \( \alpha \approx 1.26 \) for relatively large lakes and wetlands (e.g., Rosenberry et al., 2004). Therefore, when detailed temperature and vapor pressure data are unavailable, Equation (13b) may be used to estimate \( Q_e \).

For relatively small lakes and ponds having high shoreline-to-area ratios, the warm air coming from dry uplands (i.e., advection) may transfer sensible heat to the water, providing extra energy input in addition to radiation. In this case, \( Q_h \) is downward (negative) and \( Q_e \) is upward (positive), hence \( \beta \) is negative. Therefore, the actual ET flux can substantially exceed published values of potential evaporation that are based on evaporation measurements over large lakes, where radiation inputs dominate the energy balance. Some lakes and ponds are surrounded by shrubs or trees that grow in relatively moist environments, such as Salix or Populus. These riparian trees reduce shortwave radiation inputs, particularly during early spring and late fall, when the solar altitude is relatively low. More important, they block the advection of warm air from uplands and reduce the horizontal wind velocity over the pond, hence the turbulent mixing of air. In this case,
actual ET from the pond can be substantially smaller than published values of potential evaporation (Hayashi et al., 1998).

Radiation inputs have strong diurnal and seasonal variation. The variation of $Q_{h+e}$ is strongly correlated with $Q_n$ but may be modulated by the storage term ($Q_w$). In shallow ponds, where energy is stored and released on a daily cycle, ET flux remains positive during nighttime because of the energy released from storage. During daytime, on the other hand, a major portion of the incoming radiation is stored in the water, resulting in a relatively even distribution of ET flux between daytime and nighttime (Burba et al., 1999). Evaporation flux in deep lakes has a lagged seasonal response, where the flux remains high into late fall, owing to the release of energy stored in the water (Blanken et al., 2000). Transpiration by emergent plants requires transporting water from the root system to leaf surfaces, overcoming gravity and hydraulic resistance within the plant tissues. When the primary energy input is net radiation, total ET is limited by available energy incident upon a unit area regardless of the mass-transfer efficiency of leaves. In other words, transpiration mediated by emergent plants cannot be much higher than evaporation from open water surfaces. One could argue that plants with a large leaf area index can capture the sensible heat advected from uplands more effectively than smooth water surfaces. Few studies have compared the ET fluxes from open water and emergent vegetation, but it appears that average ET from vegetated water surfaces is somewhat smaller than evaporation from open water (Burba et al., 1999). Plant leaves have a much smaller capacity to store energy compared to water. As a result, the radiation inputs during daytime are almost immediately released to the atmosphere at a relatively high Bowen ratio because the temperature of leaf surfaces increases much faster than that of water. Peacock and Hess (2004) reported daytime Bowen ratios in a range of 0.5 to 2.5 over a wetland covered by Phragmites australis. They noted that ET flux was about half of reference ET (i.e., potential ET) when leaves were dry, but was close to reference ET when leaves were wet from rain. However, transpiration by individual Phragmites australis, determined by the direct measurement of sap flow through stems, can be several times higher than potential ET (Moro et al., 2004).

Evapotranspiration is a major component of the water balance, especially for closed lakes without outflow streams. Air temperature is sometimes used as an index of ET in water balance studies because it is correlated with net radiation. However, it is clear from the preceding discussion that ET is strongly influenced by wind, humidity, and energy storage in water, as well as the height and growth pattern of emergent and...
riparian plants. Therefore, a simple temperature-index method such as that given by Thornthwaite (1948) should be used with caution because it was originally intended as an index of climate, not as a predictive tool for daily or monthly ET (Dunne and Leopold, 1978). Models that use only solar radiation and temperature, however, have been shown to work surprisingly well in some lake and wetland settings (Winter et al., 1994; Rosenberry et al., 2004).

**Groundwater Exchange**

Lakes and ponds are almost always connected to groundwater (Fig. 1). Therefore, groundwater inputs and outputs always affect the water and dissolved mass balance of lakes and ponds (Winter et al., 1998; Hayashi and Rosenberry, 2002). The magnitude of groundwater flux depends on the property of substrates and the gradient of hydraulic potential according to Darcy’s law, as illustrated in an idealized example in which relatively permeable sand is underlain by impermeable bedrock (Fig. 2). The dashed line represents the water table, which, roughly speaking, is the boundary between saturated and unsaturated sand. The water level in monitoring wells indicates the hydraulic potential of groundwater, or hydraulic head, around the screen at the bottom of the well. The hydraulic head is higher at well A than at well B in Fig. 2, indicating lateral flow of groundwater towards the lake. The difference in head (i.e., potential drop) divided by the distance is the hydraulic gradient. According to Darcy’s law:

\[
Q = wdK \frac{(h_A - h_B)}{\Delta x}
\]  

(14)

![FIG. 2](image)  
**FIG. 2** Idealized example demonstrating Darcy’s law. A, B, and C are groundwater monitoring wells having short screens at the bottom. Triangles indicate the water level.
where $Q \ [L^3 \ T^{-1}]$ is the rate of groundwater flow into the lake, $w \ [L]$ is the cross sectional width perpendicular to the flow direction, $d \ [L]$ is the thickness of the saturated sediment, $K \ [L \ T^{-1}]$ is the hydraulic conductivity of the saturated sediment, $h \ [L]$ is the hydraulic head with subscripts indicating the location, and $\Delta x \ [L]$ is the distance between the two points.

In more general form, Darcy’s law is written as:

$$q = -K \ \frac{\Delta h}{\Delta l} \quad (15)$$

where $q \ [L \ T^{-1}]$ is specific discharge given by $Q/(wd)$, and $\Delta h$ is the change in head between two points separated by distance $\Delta l$ in an arbitrary direction. If the head decreases ($\Delta h < 0$) from the first point to the second point, then groundwater is flowing from the first to the second point and $q$ is positive. When the well screen is located below the water table, the hydraulic head around the screen may be different from the water table. In Fig. 2, the hydraulic head in the wells is higher than the water table, indicating that the vertical component of flow is upward. As a result, groundwater under the lakebed is flowing into the lake. Hydraulic conductivity of saturated sediments is dependent on the size and connectivity of pore spaces. Coarse materials such as sand and gravel generally have $K$ ranging from $10^{-5}$ to $10^{-2} \ m \ s^{-1}$, and fine materials such as unweathered clay commonly have $K$ as low as $10^{-11}$ to $10^{-9} \ m \ s^{-1}$ (Freeze and Cherry, 1979, p. 29). The magnitude of hydraulic gradient below lakebeds is usually within a relatively narrow range between 0 and 0.2. Therefore, hydraulic conductivity is the decisive factor controlling groundwater flux.

The volume of water contained in a unit volume of sediment is called volumetric water content. Its value is roughly equal to porosity (i.e., sediment is nearly saturated) at the water table and gradually decreases from the water table to the land surface except during major infiltration events. Water in the zone above the water table, called the vadose zone, is suspended in pores by surface tension. The vadose zone immediately above the water table is under a tension-saturated condition, where sediments remain largely or entirely saturated. This zone is called the capillary fringe. Effects of surface tension are greater in smaller pores; hence, finer sediments have a greater ability to retain water and a thicker capillary fringe. Where the water table is near the ground surface, as is the case at the riparian fringe of lakes and ponds, the existence of a capillary fringe can cause the soil to be saturated right to the ground surface so that rain cannot infiltrate and runs off to the pond instead. The hydraulic conductance of pores dramatically decreases with pore radius, and only pores containing water contribute to
the flow of water (Hillel, 1998). Therefore, hydraulic conductivity of sediments increases with water content as larger and larger pores become available for flow. Hydraulic conductivity of unsaturated sediments can be orders of magnitude smaller than that of saturated sediments, meaning that the majority of lateral groundwater flow occurs through the saturated sediments below the water table and within the capillary fringe. However, vertical flow of water in the vadose zone is important for evapotranspiration and infiltration processes.

Depth to the water table has some spatial variability within a given region, but its variability is generally much smaller than the variability of ground-surface elevation. Therefore, generally speaking, the water table is a subdued replica of the ground surface (Freeze and Cherry, 1979, p. 193). Lakes and ponds are usually located in topographic lows and hence receive groundwater discharge (Fig. 2). However, depending on the local topography and geology, a lake or pond can lose water to groundwater (i.e., groundwater recharge) or have separate areas of discharge and recharge (Born et al., 1979; Winter, 1999). The direction of groundwater exchange may also change seasonally, depending on the hydraulic gradient between the pond and the underlying groundwater (Mills and Zwarich, 1985; Winter and Rosenberry, 1995).

Theoretical studies have shown that groundwater discharge flux under lakebeds is greatest near the shore and decreases toward the lake center (Pfannkuch and Winter, 1984). In an idealized cross-section of homogeneous sediment with the uniform input of groundwater from the left side (Fig. 3), the distribution of flow lines is much denser near the shore than the lake center. In this graphical representation, each flow “tube” between flow lines carries the same amount of water. Therefore, denser flow lines indicate higher groundwater input. Pfannkuch and Winter (1984) showed that the tendency of flow lines to be crowded near the shore is greater in

**FIG. 3** Groundwater flow near a lake having half-width HW and the thickness of permeable material AM. Solid lines represent flow lines and dashed lines represent the contours of hydraulic head. (Modified after Pfannkuch and Winter [1984] with permission of Elsevier Science.)
lakes having greater width (HW) and smaller thickness of the flow domain (AM). Similar flow distribution has also been observed in numerous field studies (Shaw and Prepas, 1990). However, groundwater flux varies widely, even within small areas because heterogeneity in lakebed sediments causes hydraulic conductivity to vary substantially (Kishel and Gerla, 2002).

Geology has a major influence on groundwater exchange. Lakes located on relatively permeable materials, such as fluvial sands, tend to have high groundwater inputs. In a small (0.81 km²) lake located on glacial outwash sediments in Wisconsin, Krabbenhoft et al. (1990) reported an average groundwater input of 290 mm y⁻¹, or 27% of total water input. Rosenberry et al. (2000) also studied a small lake (0.66 km²) located on coarse glacial sediments in northern Minnesota and reported a total groundwater input of 57 L s⁻¹, or 2700 mm y⁻¹ averaged over the lake, of which about 30% was discharged at springs located on shore or near shore. These springs were relatively easily identified by the growth of marsh marigold (*Caltha palustris*). Lakes located on materials with low hydraulic conductivity, such as clay or shale, may have very little groundwater inflow or outflow so that the water balance of such lakes has no significant groundwater component.

Environmental tracer studies have shown that riparian trees may selectively use the deeper groundwater recharged under uplands, even when the shallower groundwater directly below the stream bed is readily available (Dawson and Ehleringer, 1991). Riparian trees and grasses not only intercept groundwater that would otherwise discharge into ponds but also induce drawdown of the water table around the pond perimeter (Rosenberry and Winter, 1997).

Water uptake by riparian vegetation has particularly pronounced effects on the water balance of relatively small ponds that have high ratios of perimeter (i.e., shoreline length) to area, $P/A$ [L⁻¹]. Millar (1971) monitored the water level in 75 prairie wetlands in Saskatchewan, Canada, in 1963–1969 and determined the water-level recession rates during the periods between two consecutive measurements. Out of the 146 periods, varying in length from 1 to 4 weeks, 106 periods showed a significant correlation $(p < .01)$ between the recession rate and $P/A$. Fig. 4 shows an example for the Saskatoon site, in which the recession rate increases linearly with $P/A$. Riparian trees encircling a pond block the flow of warm, dry air over the pond. Therefore, water uptake by riparian trees may be accompanied by suppression of ET from the pond. It is important to consider the complex ecohydrological linkages between ponds and riparian zones in evaluation of the water balance.
Surface Water Input and Output

Some lakes and ponds have surface water inputs through confined stream channels, while others have diffuse inputs along the shoreline (Fig. 1). The amount of input is primarily controlled by the size of the contributing area (i.e., catchment) and the amount of runoff generated from a unit area. A catchment can be determined in principle by tracing drainage divides on a topographic map using computer programs (McMaster, 2002). The catchment delineated on published maps (1:50,000 is often a convenient scale) is called the gross catchment, while the actual area that drains into a lake is called the effective catchment. Gross catchments coincide with effective catchments reasonably well on mountainous terrain, where drainage divides are clearly represented on a map. On relatively flat terrain, however, effective catchments can be substantially smaller than gross catchments because subtle topography, too small to be shown on maps, may have major effects on drainage systems (Hubbard and Linder, 1986; Hayashi et al., 2003). For example, the watershed of a lake may contain small wetlands and their associated subwatersheds that direct overland flow to the wetlands rather than to the nearby larger lake.

Flow rates in streams generally have a large degree of temporal variability. In unregulated streams without artificial reservoirs, the “baseflow” during dry periods is mostly sustained by groundwater discharge to the stream, while the high flow during storms or snowmelt consists of both groundwater and surface water inputs. Because surface water input is a

![Graph](image_url)
major component of the water balance, lake water level is strongly influenced by the amount and timing of groundwater and surface water inputs to the streams feeding the lake. Older conceptual models, developed primarily for engineering purposes, assume that most runoff during storms occurs overland when the infiltration capacity of the surface soil is exceeded by rainfall intensity. Detailed field studies have shown, however, that such a simplistic model applies only to areas covered by an impervious surface such as bedrock, urban pavement, heavily compacted soils, or frozen ground (see Dunne and Leopold, 1978, for an excellent review). In most vegetated catchments, runoff occurs as a combination of surface and subsurface processes, and overland flow is normally limited to riparian zones and lake margins where the water table is close to the ground surface.

Recent studies indicate an important role of macropores, such as decayed root holes and animal burrows, in infiltration processes (Bodhinayake and Si, 2004). Pore diameters within the soil matrix are generally less than a few millimeters and groundwater flow velocity in the soil matrix is less than a few centimeters per hour. In contrast, macropores can have diameters as large as a few centimeters and, as a result, macropore flow velocity can be several orders of magnitude greater than soil matrix flow velocity. However, the soil needs to become saturated or near saturated in order to have water in macropores. Alteration of macropores by cultivation or compaction by animals or humans may have a major influence on infiltration and subsurface flow, hence on the input of water to streams and lakes.

The amount and timing of surface and subsurface runoff are strongly dependent on soil types and geology that control the infiltration rate of rain and snowmelt. They are also dependent on the soil moisture condition because wetter soils tend to generate more runoff than do drier soils that have a larger capacity to accommodate infiltration. Vegetation has a major influence on the soil water balance through rain and snow interception and transpiration. Therefore, major land-cover change caused by natural causes, such as fire or human activities, can result in a dramatic shift in the water regime of lakes and ponds receiving water from the affected area (van der Kamp et al., 2003).

Surface water outflows from lakes and ponds are usually confined to one channel. The amount of outflow is determined by the width and depth of water, channel slope, and channel roughness, which is dependent on the type of channel material as well as presence of submergent and emergent vegetation. Flow velocity and channel cross-sectional area both increase with the depth of water, resulting in the strong dependence of volumetric
flow rate on water depth. Water depth at the outlet increases with lake stage, implying that the rate of surface outflow is greater when the amount of water stored in the lake is greater. Therefore, a negative feedback mechanism exists in the lake, through which the water level is regulated around its average position.

Storage and Basin Morphometry

The volume \( V \) and area \( A \) of lakes are usually calculated from the depth of water \( H \) by using predetermined \( V-A-H \) relations because it is difficult to monitor \( V \) and \( A \) directly. Each lake has a unique \( V-A-H \) relation reflecting the complicated morphometry of the lake basin. However, it is useful to define simple mathematical forms of \( V-A-H \) that are applicable to many single-basin lakes for examining the effects of basin morphometry on water level fluctuation. For prairie wetlands forming in isolated basins in Saskatchewan and North Dakota, Hayashi and van der Kamp (2000) proposed the following:

\[
A = s \left( \frac{H}{H_0} \right)^{2p} \tag{16}
\]

\[
V = \frac{s}{(1 + 2/p)} \frac{H^{1 + 2/p}}{H_0^{2p}} \tag{17}
\]

where \( s \) [L^2] is a scaling factor, \( p \) is a dimensionless constant, and \( H_0 \) [L] is the unit length that keeps the dimension of the equations correct. For example, \( H_0 = 1 \) m is always used if the length is measured in meters. The two parameters, \( s \) and \( p \), are specific to each basin, hence need to be determined by bathymetry survey. These equations have been successfully used for ephemeral forest pools in Massachusetts (Brooks and Hayashi, 2002).

The \( p \) takes a value of 1 for a hypothetical cone-shaped basin (Fig. 5A) and a value of 2 for a paraboloid basin, and increases with the curvature of the basin profile until it reaches \( \infty \) for a cylinder. The \( A-h \) relation is linear for a paraboloid and nonlinear for other shapes (Fig. 5B). For hypothetical basins having a depth of 1 m and area of 1000 m^2, the volume corresponding to a given depth varies with \( p \) (Fig. 5B). Real basins have complex shapes, and \( p \) gives the characteristics of the basin in an average sense. Hayashi and van der Kamp (2000) reported \( p = 1.5 \) to 6.2 for 27 prairie wetlands, and Brooks and Hayashi (2002) reported \( p = 0.6 \) to 2.2 for 33 forest pools. Ponds with larger \( p \) have a more concave basin shape to store a larger amount of water for a given area; they also have smaller rates of
FIG. 5 Idealized symmetric basins (A) and their volume-area-depth relation (B).

area shrinkage for a given water level drop. These are important factors regarding the ecohydrological response of ponds to changes in inputs or outputs of water. For example, the zonation patterns of emergent vegetation are controlled by the changes in the depth of water and the location of the shoreline (van der Valk, 2000; also see Chapter 11).
Climate Effects

The average water level in ponds and lakes, and the water level changes around that average, reflect a dynamic balance among various water inputs and outputs (Fig. 1). The riparian vegetation around the water body responds to the degree of hydrological disturbance that is represented by the short-term and long-term changes of water level. In fact, one can often deduce a considerable amount of information about the water-level regime of a pond or lake by inspection of the vegetation in and around it.

The average water level is determined by landscape setting, hydrogeological setting, and the climate. In humid climates, where precipitation exceeds ET annually, most ponds and lakes have a surface outflow that controls the average water level and restricts water level changes to a fairly narrow range. High water levels caused by storms or snowmelt will tend to dissipate rather quickly. However, some ponds and lakes located on permeable materials are drained mainly by groundwater outflow. Since groundwater outflow is less sensitive to lake level fluctuations, changes in water inputs in these ponds and lakes lead to larger and longer-lasting changes of water level.

In semi-arid and arid climates, where ET exceeds precipitation on an annual basis, lakes and ponds can be hydrologically closed with no outflow. ET is then the dominant or only form of water output. Unlike water outflow by surface or groundwater, ET is not directly controlled by water level, although total lake ET is dependent on the lake area that expands and shrinks depending on water level. Therefore, large, closed water bodies may have only weak feedback mechanisms to stabilize the water level if the lake area does not change appreciably with changes of water level. Such lakes can be subject to large, multiyear water level changes. They may experience large sudden rises of water level as a result of high runoff, but the subsequent downward recession is slow, being limited by the difference between precipitation and evaporation. The water level regime of large closed lakes can be difficult to simulate or predict because relatively small changes in evaporation and precipitation, or of inputs such as changes of runoff due to landuse change, can lead to large cumulative changes of water level over the long term. As an example, Devil's Lake in North Dakota is a closed lake having a catchment of 8600 km² (Todhunter and Rundquist, 2004). The lake level increased by about 7 m between 1992 and 2001, most likely because of a shift of climate from dry to wet, although the effects of agricultural land conversion and drainage are also debated (Todhunter and Rundquist, 2004).
CASE STUDY: NORTHERN PRAIRIE WETLANDS

Overview of Prairie Wetlands

The northern prairie region of North America is characterized by undulating topography with numerous depressions. Prairie wetlands forming in these depressions are the focal point of aquatic ecosystems in the region. Dynamic water regimes of the prairie wetlands provide excellent case study examples demonstrating the effects of hydrological processes on the water levels. The studies presented below were conducted in St. Denis National Wildlife Area located 40 km east of Saskatoon, Saskatchewan, Canada. Over 100 wetlands and numerous smaller depressions occur on the 4 km² study site. The land use consists of the conventional rotation of cereal crops and oil seeds, tame grasses, and native prairie grasses.

Prairie wetlands usually have no permanent surface inflow or outflow streams and are regarded as hydrologically closed basins. The water balance is strongly influenced by the cold semi-arid climate, where annual evaporation from open water (600–800 mm) exceeds annual precipitation (300–400 mm). Winter snowfall accumulates up to 150 mm of water equivalent and the soil freezes to depths greater than 1 m (van der Kamp et al., 2003). Glacial tills underlying the surface soils are rich in clay and have relatively low hydraulic conductivity, limiting the rates of groundwater flow (Hayashi et al., 1998).

Major inputs of water are direct precipitation onto wetlands and the lateral transfer of snow-derived water from adjacent lands in the form of wind-driven snowdrift and snowmelt runoff over frozen ground. Summer runoff is rare because of the dry climate, and groundwater input is minor because of the low hydraulic conductivity of glacial till. Outputs are evapotranspiration and infiltration of surface water to groundwater since there is no outflow stream.

The water level is generally highest in early spring after snowmelt runoff is received and gradually declines in summer, primarily because of evaporation and infiltration exceeding the inputs of rain and occasional storm runoff (Winter, 1989). Many prairie wetlands hold surface water only for a few weeks to months. The duration of surface water, or hydroperiod, is a critical habitat parameter for waterfowl and other species dependent on water (Swanson and Duebbert, 1989). The water level change in a typical prairie wetland can be schematically represented by a triangle (Fig. 6) where the magnitude of spring rise and the slope of summer recession determine the duration of surface water. Therefore, the ecology of prairie wetlands is strongly dependent on the factors controlling spring rise and summer recession.
Effects of Upland Vegetation and Land Use

The magnitude of spring rise depends on the volume of snow drift and snowmelt runoff into the wetland. Snow drift is influenced by many factors, but microtopography and vegetative cover on the upland seem to be the most important (van der Kamp et al., 2003). Tall stubble and perennial grasses retain snow on the upland, while resuspension and redistribution of snow is relatively unrestricted on cultivated fields with little stubble. Riparian trees around wetlands also function as effective snow accumulators (Hayashi et al., 1998). In the absence of snow-trapping vegetation, snow accumulates preferentially in depressions.

The volume of snowmelt runoff is determined by the size of the catchment and the amount of runoff generated over a unit area. The latter depends on the amount of snow on the ground and the infiltration capacity of frozen soil, which in turn depends on soil water content, particle size, and structure (Gray et al., 2001). Conditions favorable for large amounts of runoff are wet soil at the time of freeze up, high clay content, and the absence of soil macropores, such as decayed root holes and animal burrows (van der Kamp et al., 2003). Macropores can easily be destroyed by cultivation. Therefore, development of macropores is enhanced under undisturbed vegetation compared to cultivated fields.

Long-term records of water level at Wetland 92 (Fig. 7A) and Wetland 109 (Fig. 7B) show that the two wetlands had similar water regimes until 1986. The two wetlands have similar size, and the uplands around these wetlands were subjected to dryland cultivation until 1983, when the uplands around 92 were converted to a dense nesting cover of Bromus inermis for the purpose of improving wildlife habitat. This dense nesting cover has not been disturbed by grazing, mowing, or burning (van der Kamp

**FIG. 6** Annual cycle of water level in typical prairie wetlands.
et al., 2003). From 1987 onward, Wetland 92 dried out and has remained almost entirely dry ever since, while the water regime in Wetland 109 did not show a major change. Other small wetlands within the grassed area behaved similarly to Wetland 92 (van der Kamp et al., 1999). Detailed studies of hydrological processes indicated that two factors operate together and lead to the dramatic shift in wetland water regime: (1) tall permanent grass cover is effective in trapping snow so that the wind-driven transport of snow into wetlands is reduced; and (2) the undisturbed grass cover leads to the development of a macropore network in the topsoil, which markedly increases the infiltrability of the soil, even when it is frozen.

FIG. 7  Water depth records for (A) Wetland 92, (B) Wetland 109, and (C) Wetland 50. The uplands around Wetland 92 were converted to permanent grass in 1983.
The macropore network takes several years to develop after introduction of the grass, as indicated by the delayed response of water level in 1987 to the cultivated-to-grass conversion that took place in 1983.

**Size and Permanence of Wetlands**

In addition to the triangular annual cycles (Fig. 6), water levels in wetlands have interannual and interdecadal cycles related to the variability of climatic conditions. For example, relatively dry conditions in the late 1980s to early 1990s resulted in low water levels, which recovered during the relatively wet period of the mid 1990s (Fig. 7B). Winter and Rosenberry (1998) found similar patterns in the water levels of prairie wetlands in North Dakota and showed the correlation between water levels and a climatic index. Wetlands with seasonal ponds, such as Wetland 109, respond immediately to reduced water inputs during dry periods, resulting in a very short duration of surface water (Fig. 7B). In contrast, wetlands with more permanent ponds, such as Wetland 50, keep their surface water during multiple years of low water inputs (Fig. 7C).

The area of Wetland 109 is approximately 0.24 ha and its catchment is 2.4 ha (Hayashi et al., 1998), meaning that the catchment-to-wetland area ratio is 10. The area of Wetland 50 is 3.4 ha and its catchment-to-wetland area ratio is 16 (Su et al., 2000). Therefore, if the upland areas generate constant amounts of runoff on average, Wetland 50 receives 60% more spring runoff per unit area than Wetland 109. This partially explains why the former has a more permanent pond than the latter. The permanence of a pond is also related to its $P/A$. The rate of water level recession in summer is higher for those ponds with larger $P/A$ than those with smaller $P/A$ (Fig. 4). Therefore, smaller ponds in smaller wetlands have a larger $P/A$ and tend to have shorter durations. Approximate $P/A$ at peak water level is 0.02 m$^{-1}$ for the pond in Wetland 50 and 0.07 m$^{-1}$ for the pond in Wetland 109. Fig. 4 suggests that summer recession rates of the water level in Wetland 109 could be nearly 50% higher than that in Wetland 50, and the water level data in Fig. 7 show that this is indeed the case, with the water level recession in Wetland 109 always being greater than the corresponding recession in Wetland 50.

Many of the wetlands with permanent and semi-permanent ponds have high salinity, indicating the accumulation of dissolved salts transported to wetlands by groundwater and surface water. However, except for the shoreline-related seepage discussed below, the flow rate of groundwater is typically much less than 50 mm y$^{-1}$ (van der Kamp and Hayashi, 1998), which is too low to have measurable effects on the water balance. Therefore, the groundwater input generally has little effect on the water
balance. However, the groundwater input is important for the mass balance of dissolved chemical species, such as sulfates, because the concentration of these species in groundwater is several orders of magnitude higher than that in rain or snowmelt water.

**Riparian Vegetation**

The shoreline-related loss of water as indicated in Fig. 4 is linked to water uptake in the riparian zone around the wetland, where the water table is close to the ground surface, allowing plants to transpire at a rate approaching potential ET. Fig. 8 shows the elevation of the water surface in Wetland 109 and hydraulic head in two monitoring wells under the riparian zone.

**FIG. 8** (A) Daily precipitation, (B) air temperature, and (C) pond and piezometer water level at Wetland 109.
The relative position of the pond, monitoring wells, and riparian zone are schematically shown in the insert. The distance between W1 and W2 is approximately 7 m. The distance between the edge of the pond and W2 ranged between 5 and 13 m during this period. Hydraulic head decreased from the pond to W1, then to W2, indicating that the groundwater flow direction was away from the pond towards the edge of the wetland. Hydraulic head data show peaks around 10:00 to 11:00 and troughs around 19:00 to 22:00. It is clear that daytime drawdown of the groundwater level is caused by water uptake by plants. The nighttime recovery most likely represents the replenishment of water by the lateral flow of groundwater from the pond to the riparian zone. Parsons et al. (2004) reported that 60% to 70% of water loss from the pond in Wetland 109 during the summer months was caused by the groundwater flow induced by the riparian vegetation uptake, while the remainder was evapotranspiration within the pond area.

The clear diurnal pattern at W1 and W2 disappeared after September 17, and the slope of the pond water recession decreased, indicating that the water uptake by riparian plants had stopped. Air temperature dropped to 0.6°C on September 11 (Fig. 8B), suggesting a possibility of the first frost because the temperature of exposed leaves can be substantially lower than air temperature because of radiation cooling (Inouye, 2000). We did not document the phenology of trees, but it is likely that leaf senescence started sometime between September 11 and 17.

The water levels in the pond and wells rose rapidly on September 5 and 6 (Fig. 8C), responding to rainfall (Fig. 8A). The pond level rose by 43 mm, where total rainfall for this event was 36 mm, suggesting an input of 7 mm of “extra” water. This input was probably surface runoff generated in the riparian zone, where the water table is near the ground surface and the ground is almost entirely saturated because of a relatively thick capillary fringe in the clay-rich soil. The rainfall caused a rapid rise of the water table, leading to fully saturated conditions in the ground that restricted the infiltration of rainfall. Gerla (1992) and Winter and Rosenberry (1995) made similar observations at prairie wetlands in North Dakota.

CONCLUSIONS

Water-level changes in ponds and lakes occur as a result of the water input exceeding output or vice versa. Because inputs and outputs are controlled by hydrological processes, understanding the water-level changes and resulting ecological responses requires understanding the individual processes. It is
particularly important to realize the intimate link between lakes and their catchments. Disturbance in the catchment, such as major land use change, can cause a dramatic change in hydrological processes, which ultimately affects the lake water level. This was clearly demonstrated in the case study of prairie wetlands, where grassing the uplands resulted in the drying out of wetlands. Climate changes also have major effects on water level. When the hydrological processes and their response to land use and climate are understood, it is reasonably straightforward to simulate the water level in a particular lake or pond using the water balance equation (e.g., Poiani et al., 1996; Su et al., 2000). However, hydrological processes are poorly understood in many cases, resulting in large uncertainty in model predictions. Ecohydrological linkage between plants and water presents a fruitful opportunity for collaboration between ecologists and hydrologists. The role of riparian trees in evapotranspiration and groundwater exchange, for example, is an important but relatively poorly understood process. It is hoped that collaborative research on ecohydrology will enable us to observe hydrological processes and ecological responses simultaneously and to develop coupled models for the prediction of ecosystem responses to land use and climate changes.

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APPENDIX: LIST OF SYMBOLS

\[a\] constant in wind function (Equation 6) [L T^{-1}]
\[A\] area of lake or pond [L^2]
\[b\] constant in wind function (Equation 6)
\[C_p\] specific heat of air [L^2 T^{-2} K^{-1}]
\[C_w\] specific heat of water [L^2 T^{-2} K^{-1}]
\[d\] thickness of saturated sediments [L]
\[e_a\] vapor pressure of air [M L^{-1} T^{-2}]
\[e_s\] vapor pressure at the water surface [M L^{-1} T^{-2}]
\[E\] evaporation rate [L T^{-1}]
\[f(u)\] wind function [L T^{-1}]
\[h\] lake water level [L]


$\text{h}_A$  groundwater hydraulic head [L]

$H$  depth of lake or pond [L]

$H_0$  unit depth [L]

$K$  hydraulic conductivity [L T$^{-1}$]

$L$  length

$L_e$  latent heat of vaporization [L$^2$ T$^{-2}$]

$M$  mass

$p$  shape constant in Equation (16)

$p_b$  barometric pressure [M L$^{-1}$ T$^{-2}$]

$P$  perimeter of lake or pond [L]

$q$  groundwater specific discharge [L T$^{-1}$]

$Q$  groundwater flow rate [L$^3$ T$^{-1}$]

$Q_a$  net energy input by stream and groundwater per area [M T$^{-3}$]

$Q_e$  latent heat transfer [M T$^{-3}$]

$Q_g$  energy flux into lake bottom sediment [M T$^{-3}$]

$Q_b$  sensible heat transfer [M T$^{-3}$]

$Q_{b+e}$  sum of latent and sensible heat [M T$^{-3}$]

$Q_{in}$  total water input [L$^3$ T$^{-1}$]

$Q_{LW \uparrow}$  outgoing long-wave radiation [M T$^{-3}$]

$Q_n$  net radiation [M T$^{-3}$]

$Q_{out}$  total water output [L$^3$ T$^{-1}$]

$Q_w$  rate of energy storage change per lake area [M T$^{-3}$]

$s$  scaling constant in Equation (16) [L$^2$]

$T$  time

$T_a$  air temperature [K]

$T_w$  average lake temperature [K]

$T_s$  temperature of surface [K]

$u$  wind speed [L T$^{-1}$]

$V$  lake or pond volume [L$^3$]

$w$  cross-sectional width of groundwater flow [L]

$z_0$  roughness length [L]

$\alpha$  Priestley–Taylor coefficient

$\beta$  Bowen ratio

$\Delta$  slope of vapor pressure-temperature curve [M L$^{-1}$ T$^{-2}$ K$^{-1}$]

$\varepsilon$  emissivity

$\gamma$  psychrometric constant [M L$^{-1}$ T$^{-2}$ K$^{-1}$]

$\rho_a$  density of air [M L$^{-3}$]

$\rho_w$  density of water [M L$^{-3}$]

$\sigma$  Stefan–Boltzmann constant [M T$^{-3}$ K$^{-4}$]
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