Quantifying wetland–aquifer interactions in a humid subtropical climate region: An integrated approach

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1. Introduction

The hydrologic regime of a wetland is the result of a complex interplay between climate and wetland characteristics (Winter, 1998). To evaluate the capacity of a wetland to degrade contaminants, it is necessary to conduct long term studies that include observations and modeling at different scales in time and space. Quantifying the responses of an aquifer–wetland system to climatic variables is useful to understand a wide range of natural processes such as the intensity and duration of groundwater inputs and regional groundwater flow contributes little to the overall water balance. Observations and modeling at different scales in time and space are seasonal and readily respond to rainfall events. The wetland water balance is dominated by local groundwater inputs and regional groundwater flow contributes little to the overall water balance.

Winter and Harvey (1998) concluded that for accurate quantification of GW–SW interactions at the Norman landfill research site in Oklahoma, USA. Our integrated approach involved model evaluation by means of the following independent measurements: (a) groundwater inflow calculation using stable isotopes of oxygen and hydrogen ($^{16}$O, $^{18}$O, $^2$H, $^3$H); (b) seepage flux measurements in the wetland hyporheic sediment; and (c) pan evaporation measurements on land and in the wetland. The integrated approach was useful for identifying the dominant hydrological processes at the site, including recharge and subsurface flows. Simulated recharge compared well with estimates obtained using isotopic methods from previous studies and allowed us to identify specific annual signatures of this important process during the period of study (1997–2007). Similarly, observations of groundwater inflow and outflow rates to and from the wetland using seepage meters and isotope methods were found to be in good agreement with simulation results. Results indicate that subsurface flow components in the system are seasonal and readily respond to rainfall events. The wetland water balance is dominated by local groundwater inputs and regional groundwater flow contributes little to the overall water balance.

Wetlands are widely recognized as sentinels of global climate change. Long-term monitoring data combined with process-based modeling has the potential to shed light on key processes and how they change over time. This paper reports the development and application of a simple water balance model based on long-term climate, soil, vegetation and hydrological dynamics to quantify groundwater–surface water (GW–SW) interactions. The integrated approach involved model evaluation by means of the following independent measurements: (a) groundwater inflow calculation using stable isotopes of oxygen and hydrogen ($^{16}$O, $^{18}$O, $^2$H, $^3$H); (b) seepage flux measurements in the wetland hyporheic sediment; and (c) pan evaporation measurements on land and in the wetland. The integrated approach was useful for identifying the dominant hydrological processes at the site, including recharge and subsurface flows. Simulated recharge compared well with estimates obtained using isotopic methods from previous studies and allowed us to identify specific annual signatures of this important process during the period of study (1997–2007). Similarly, observations of groundwater inflow and outflow rates to and from the wetland using seepage meters and isotope methods were found to be in good agreement with simulation results. Results indicate that subsurface flow components in the system are seasonal and readily respond to rainfall events. The wetland water balance is dominated by local groundwater inputs and regional groundwater flow contributes little to the overall water balance.

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The landfill was not lined and a leachate plume extends downhill and capped with locally obtained clay, silt, and sand materials. The site received wastes from 1922 to 1985, at which time it was closed and GW–SW fluxes have been measured at the site (Cozzarelli et al., 2011; Masoner et al., 2008; Báez-Cazull et al., 2011). The site is located in Cleveland County, Oklahoma, USA (Fig. 1) and is on the south side of the Canadian River. The climate of the site is between humid subtropical and semi-arid, with an average annual temperature of 16 °C, and average maximum and minimum temperatures of 23 °C and 9 °C respectively during the period 1957–2009. The hottest month is July with an average temperature of 28 °C. The coldest month is January with an average temperature of 3.5 °C and the average annual precipitation is 88 cm.

As reported by Cozzarelli et al. (2011), the Canadian River alluvium is 10–12 m thick and predominantly composed of fine- to medium-grained sand beds with interbedded, discontinuous layers of clayey silt and gravel at depths between 3 and 5 m below the ground surface. Measured hydraulic conductivity of the unconfined alluvial aquifer ranges from 8.4 × 10⁻⁷ to 2.8 × 10⁻⁴ m/s with a median value of 6.6 × 10⁻⁴ m/s (Scholl and Christenson, 1998). The aquifer is underlain by the Hennessy Group, a shale and mudstone confining unit. A potentiometric surface map of the area made in 1995 (10 years after the landfill was capped) shows regional groundwater flow toward the Canadian River with a hydraulic gradient of about 1.4 m/km south of the landfill (Scholl and Christenson, 1998).

The present study focused on local interactions between groundwater and the wetland. The specific study area includes the area of the wetland (actual area inundated) of 8800 m² and the riparian catchment area of 38,755 m² shown in Fig. 1. The wetland is approximately 700 m long and 15–25 m wide and is 50–100 m from the southern toe of the landfill. The wetland is situated in a previous location of the main river channel and is aligned perpendicular to the groundwater flow path. Water level in the wetland is an expression of the regional water table with the wetland pool elevation usually being lower than the up-gradient groundwater elevation. Dry periods have been observed during summer when the water table drops below the wetland bottom. It has been reported that the wetland has no apparent surface–water sources and is mainly fed by groundwater discharge and precipitation (Cozzarelli et al., 2011; Masoner et al., 2008; Báez-Cazull et al., 2007). The wetland system is a shallow stream, with ponded areas or wetlands caused by beaver dams. A road that intersects the upstream section of the wetland (Fig. 1) acts as a dam that limits surface flow from upstream ponded areas. At the down gradient end of the wetland there is a beaver dam, isolating the wetland to pools downstream. Seepage measurements showed that the groundwater connectivity along the direction of the wetland is insignificant. Cozzarelli et al. (2011) reported that the geological conditions at the site have created a leachate plume that migrates underneath the wetland toward the Canadian River and also interacts with the wetland (see Fig. 2 in Cozzarelli et al. 2011). Therefore in this study it is assumed that groundwater flow and precipitation are the dominant inputs into the wetland.

The riparian area near the landfill is densely vegetated with shallow-rooted vegetation, also found in the wetland area and deep-rooted vegetation dominated by at least three species of phreatophytes namely: willow, cottonwood, Eastern red cedar, and salt cedar (Burgess, 2004; Scholl et al., 2005). The wetland vegetation is composed of a mixture of native and introduced species including: common reed, western ragweed, Bermuda grass, John-son grass, bandedflower, Ravenna grass, giant cane, sandbar willow and black willow (Burgess, 2004; Masoner et al., 2008; Zume et al., 2006).

2. Site description

To test the bucket model we use field data collected at the Norman Landfill Research Site. The site is ideal for this research in that extensive geochemical data, evaporation rates, piezometric data and GW–SW fluxes have been measured at the site (Cozzarelli et al., 2011; Masoner and Stannard, 2010). The site is located in Cleveland County, Oklahoma, USA (Fig. 1) and is on the south side of the City of Norman and includes a former municipal landfill that received wastes from 1922 to 1985, at which time it was closed and capped with locally obtained clay, silt, and sand materials. The landfill was not lined and a leachate plume extends downgradient from the landfill in the direction of regional groundwater flow (Becker, 2001), which is towards the wetland and the Canadian River. The climate of the site is between humid subtropical and semi-arid, with an average annual temperature of 16 °C, and average maximum and minimum temperatures of 23 °C and 9 °C respectively during the period 1957–2009. The hottest month is July with an average temperature of 28 °C. The coldest month is January with an average temperature of 3.5 °C and the average annual precipitation is 88 cm.

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3. Materials and methods

Readily available meteorological data and initial water levels were used as inputs to the model described in the next section. Measured soil properties were used to constrain parameters. Evaporation pan data were used for comparison of this flux with...
simulated results. Isotopic analyses and seepage measurements were used to obtain groundwater flows and to compare with simulated groundwater inflow to and outflow from the wetland.

Meteorological information was retrieved from the Norman station of the Oklahoma MESONET (www.mesonet.org). For precipitation, the data were retrieved from the Norman station and four different MESONET meteorological stations around the area (Oklahoma City East, Minco, Washington, and Shawnee) and the USGS Gauge Station (07229053) for the Canadian River Tributary at Norman. Daily precipitation data sets from each of these six stations were used to calculate an average of daily precipitation for the site using Thiessen polygons. Measured values of porosity ranged from 0.34 to 0.44 with an average value of 0.40 from samples in the field. A study of spatial variation of hydraulic conductivity using slug tests (Scholl and Christenson, 1998) reported a range from $8.4 \times 10^{-7}$ to $2.8 \times 10^{-4}$ m/s with a median value of $6.6 \times 10^{-5}$ m/s for the alluvium aquifer in the research site area.

Water table depths from 1997 to 2006 were obtained from a well located 80 m north of the wetland (Plume well in Fig. 1). The water table in the alluvial aquifer varies seasonally, generally ranging from 1.3 to 3.2 m below ground surface (elevation of 331.04–329.11 m) in the area between the landfill and the wetland. Water table depth in the well was recorded at every hour for the period May 1997–December 2006 (http://ok.water.usgs.gov/projects/norlan/other.html). Hourly data were used to obtain average daily water table depth values.

Evaporation rates were measured with a standard Class A evaporation pan on land and a modified Class A floating evaporation pan in open water of the wetland (Fig. 2a) for the period of February 14 to August 31, 2005. Further details of the pan measurements including details of pan construction and operation can be found in Masoner et al. (2008). Water level data are available at the USGS gauge height site 07229053 of the Canadian River Tributary at Norman, OK, where the elevation reference level of the gauge is 328.97 m. Location of the gage is depicted in Fig. 1 (weather station). Water level data for the wetland used in this study were for the period of May 1997 to December 2006. The measured height of the water in the wetland varied from 0.04 to 1.3 m (elevations from 329.00 to 330.30 m).

Isotopic analyses of $^{18}$O and $^2$H were conducted from March 20, 2007 to September 15, 2008. Groundwater samples were obtained from two different well transects located across the wetland constructed in 1997 as part of the Toxic Substances Hydrology Program of the U.S. Geological Survey (http://ok.water.usgs.gov/).
Numerous figures exist in the Norman Landfill literature showing that the two well transects are parallel to the groundwater flow direction (Cozzarelli et al., 2011; Becker, 2001; Báez-Cazull et al., 2007, 2008; Christenson and Cozzarelli, 2003; Eganhouse et al., 2001). Based upon continuous research at the site it was observed that the groundwater flow direction is generally southwest between the landfill and the wetland. Transects 148 and 153 are composed of 7 and 8 wells respectively (Fig. 1) and were designed to follow the groundwater flow path. Wells in the two transects were named sequentially following the groundwater flow direction. For example, 148B is located up-gradient and 148H is located down-gradient from the wetland. Groundwater at depths of less than 5 m was sampled at time intervals ranging from 9 to 55 days, but usually once per month from all the wells of the 148 and 153 transects. Wetland samples were taken by boat between well transects 148 and 153. Depth averaged samples were obtained manually by withdrawing water near the top, middle, and bottom depths with a 100 ml syringe. Average wetland depth was around 1.06 m during the sampling period. Precipitation samples for stable isotope analyses were collected biweekly in a funnel/bottle collector containing a 1-cm layer of mineral oil to prevent evaporation. The funnel was 17.4 cm in diameter, inserted through a stopper in a 2-l collection bottle (Scholl et al., 2005; Jaeschke et al., 2011). Location of the precipitation sample collector is depicted in Fig. 3. All isotope samples were analyzed in the USGS Reston Stable Isotope Laboratory. Details of oxygen and hydrogen isotopic analysis can be found in Révész and Coplen (2008a,b).

4. Model development and application

4.1. Conceptual framework

The conceptual framework of the water balance model is based on findings from previous potentiometric surface measurements and geochemical data for the wetland and shallow groundwater, which indicated that, along most of the wetland, leachate-containing groundwater discharges to the wetland along the northeast bank and wetland water recharges the aquifer along the southwest bank (Báez-Cazull et al., 2007; Cozzarelli et al., 2000; Grossman et al., 2002; Schottmann et al., 1999; Scholl et al., 2005; Scholl et al., 2006). In this system, the water table is shallow and the wetland is in direct connection with it as a surface expression of the water table (Winter et al., 1995; Semeniuk and Semeniuk, 1995).

The conceptual model of the groundwater–wetland system is illustrated in Fig. 3. The system has a defined topographic boundary and is subdivided into the riparian catchment area and the wetland area. The riparian catchment funnels water toward the wetland. Water table elevation in the riparian catchment changes continually in response to the following processes: precipitation (\(p\)), bare soil evaporation (\(e_{bs}\)) in areas devoid of vegetation, and transpiration from vegetation in the unsaturated (\(e_{us}\)) and saturated zones (\(e_{sat}\)) in areas covered with vegetation. Three more processes influence the water table depth, namely recharge to groundwater (\(Re\)) from the unsaturated zone, subsurface flow (\(Q_{reg}\)) that feeds the regional groundwater flow to the Canadian River, and local groundwater flow from the riparian catchment to the wetland area (\(q_{local}\)).

Transient characteristics of water depth in the wetland are influenced by precipitation (\(p\)), ET (\(e_{WL}\)), local groundwater flow received from the upland area (\(Q_{local}\)), and local subsurface outflow to the aquifer from the wetland area (\(q_{out}\)). Since the wetland is a surface expression of the water table and is part of the flow system connected to the main river, a regional subsurface flow (\(Q_{reg}\)) that feeds the regional groundwater flow will also affect its water level. The development of the water balance model is described in the following section.

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Fig. 2. Photographs of (a) Floating pan deployed in the wetland, modified from Masoner et al. (2008) and (b) Seepage meters. Photographs were taken in the field by Jason R. Masoner.

Fig. 3. Conceptual framework of the groundwater–wetland system.
4.2. Water balance model

The water balance model was based on a leaky bucket model (i.e., a bucket model with outflow) approach (Farmer et al., 2003; Jothityangkoon et al., 2001; Krasnowstein and Oldham, 2004). The bucket model is similar to the interactive groundwater approach of Krasnowstein and Oldham (2004) but differs in that our model incorporates differentiation between bare soil evaporation, transpiration and ET in the riparian and wetland areas. In our model we conceptualized two buckets: the riparian catchment represents one bucket composed of an unsaturated zone and a saturated zone and the wetland area represents another bucket. Conceptually, the riparian catchment is a leaky bucket (with soil properties such as porosity and field capacity) that receives water from precipitation and loses water through evaporation, ET and transpiration. In the riparian catchment, precipitation is assumed to readily infiltrate into the unsaturated layer. Surface-water discharge to the wetland has not been observed at the site, thus the model assumes that there is no overland runoff for the daily time-step used in the model. It is recognized that ponding or runoff play an important role in similar systems, the advantage of the bucket model is that these processes can be added to the model by either adding a sink term in the soil moisture bucket or by adding a bucket that considers a wetting front approach such as Struthers et al. (2006) if considered necessary. When soil moisture meets its limited capacity to hold water, the excess is sent to the groundwater system as recharge. In the groundwater system, recharge is an input, and the outputs are composed of transpiration and subsurface flow to both the regional and the local systems. When the water level in the groundwater system is higher than in the wetland, water is sent to the wetland via local subsurface flow. In this model it is assumed that only the top of the aquifer contributes to the wetland, however the regional flow influences fluctuations of the water table depth. The wetland area represents another leaky bucket (without soil properties) that receives water from precipitation and a local groundwater flow from the riparian bucket and loses water via ET. The wetland area outputs are regional groundwater flow and local subsurface outflow and occur when the water level up-gradient is higher than the water level down-gradient.

4.2.1. Water balance in the riparian catchment

The riparian catchment is assumed to receive water from recharge from the unsaturated zone (Re) and lose water through transpiration from phreatophyte vegetation (eVs), regional groundwater flow qreg (t) and local subsurface flow to the wetland (qlocal). The water balance in the riparian catchment is represented by the following equation:

\[
s(t + \Delta t) = s(t) + \frac{Re(t)}{\varphi(1 - \theta_c)} - \left(\frac{eVs(t) + qlocal(t) + Qreg(t)}{\varphi(1 - \theta_c)}\right) \Delta t \tag{1}
\]

where \(s\) is the water storage in the riparian catchment per unit area, eVs is the transpiration from the groundwater or saturated zone by phreatophyte vegetation (cm/day), qlocal is the groundwater local subsurface flow component (cm/day), Qreg is the regional groundwater subsurface flow (cm/day), Re is the recharge from the unsaturated zone to the groundwater or saturated zone, \(\varphi\) is the porosity of the soil, and \(\theta_c\) is the field capacity. Here \(\theta_c\) is a water saturation fraction of the pore spaces after infiltration. The field capacity defines the maximum soil water moisture that will remain in the soil after infiltration. This is different from the soil science definition that requires 2 days of drainage but has been successfully used before by Farmer et al. (2003), Muneepeerakul et al. (2008) and Krasnowstein and Oldham (2004). The amount of water that recharges the groundwater depends on soil moisture, evaporation and transpiration in the unsaturated zone as explained below.

4.2.1.1. Soil moisture, evaporation and transpiration in the unsaturated zone

The unsaturated zone is defined as the soil column above the water table up to the surface (assuming that the capillary fringe is small enough to be negligible and thus no air entry point is situated). Precipitation falls on the riparian catchment and infiltrates to the unsaturated zone, where water is lost from the unsaturated zone as groundwater recharge (Re), bare soil evaporation (eVs), and transpiration from the area covered by vegetation (eIUS). The governing equations for soil moisture (\(\theta\)) are as follows (Muneepeerakul et al., 2008):

\[
\theta(t + \Delta t) = \theta(t) + \left(\frac{IUS(t) - eVs(t) - eIUS(t)}{\varphi(y)}\right) \Delta t \text{ if } IUS(t) \Delta t \leq WSC(\theta, y) \tag{2}
\]

\[
\theta(t + \Delta t) = \theta(t) + \left(\frac{eVs(t) - eIUS(t)}{\varphi(y)}\right) \Delta t \text{ if } IUS(t) \Delta t > WSC(\theta, y) \tag{3}
\]

where \(\theta\) is the moisture content of the soil (expressed as a fraction of the pore space), WSC is the water storage capacity of the unsaturated zone, eIUS is the transpiration from the unsaturated zone (cm/day), \(\varphi\) is the porosity of the soil, \(\Delta t\) is the time step of the model simulation (1 day), and \(y\) is the simulated water table depth in cm. The water storage capacity of the unsaturated zone (WSC) is given by (Muneepeerakul et al., 2008):

\[
WSC(\theta, y) = \varphi(y) [\theta_c - \theta(t)] \tag{4}
\]

The term \(y^*\) in Eq. (3) is the water table depth (distance from the ground surface to the water table) immediately after infiltration:

\[
y^*(t) = y(t) - \frac{Re(t, \theta)}{\varphi(1 - \theta_c)} \tag{5}
\]

where Re is the recharge. Discharge is defined as the amount of water reaching the groundwater and is given by:

\[
Re(\theta, y) = \left\{\begin{array}{lr}
IUS(t) \Delta t - WSC(\theta, y) & \text{if } IUS(t) \geq WSC(\theta, y) \\
0 & \text{if } IUS(t) < WSC(\theta, y)
\end{array}\right. \tag{6}
\]

where \(IUS\) is the amount of water infiltrating the unsaturated zone in cm/day.

In the riparian catchment, precipitation is assumed to readily infiltrate into the unsaturated layer and is given by:

\[
IUS(t) = \varphi(t) \tag{7}
\]

where \(\varphi\) is the measured precipitation in cm/day (see Meteorological data subsection). This means that if the infiltration rate \(IUS\) is lower than the capacity of the soil to continue to hold water (WSC), soil moisture will increase and infiltration from precipitation remains stored in the unsaturated zone. When infiltration is higher than the water holding capacity, soil moisture in the unsaturated zone is left at the maximum moisture capacity. The maximum moisture capacity in this study is defined as the field capacity expressed as a fraction of the pore space. As explained in the Site Description section, the Canadian River alluvium is predominantly fine- to medium-grained sand beds. There are interbedded, discontinuous layers of clayey silt and gravel at depths between 3 and 5 m below the ground surface in the area of study, which are below the water table depth (minimum 1.3 and maximum 3.2 m below ground surface). Therefore field capacity can be considered homogeneous and the unsaturated zone composed of sandy soil, where the assumption that there is no ponding or runoff in 1 day time step is reasonable.

After infiltration, water that remained in the unsaturated zone is subjected to bare soil evaporation and transpiration from phreatophytes that use water from the saturated and unsaturated zones. At the Norman site, land in the riparian area is covered with
persistent deep-rooted vegetation and some areas are either de-
void of vegetation or with shallow-rooted vegetation. Therefore,
we defined bare soil evaporation as evaporation from the percent-
age of soil devoid of vegetation and/or covered with shallow-
rooted vegetation.

For long-term simulations, the model uses reference evapo-
transpiration rates. Pan evaporation rates can be used as reference
ET rates; however they were measured for a short period of time
(February to August 2005). Therefore the Penman–Monteith (here-
inafter P–M) equation (Allen, 1986; Allen et al., 1998; Monteith
and Unsworth, 1990) was used to obtain reference ET rates for
the 10-year simulation period. Many different equations can be
used for calculating ET losses from wetlands (Rosenberry et al.,
2004). We used the P–M equation to compute the reference ET rate
$E_{\text{PM}}(t)$ as it is known to perform well in different settings (Drexler
et al., 2004; Rosenberry et al., 2004; Summer and Jacobs, 2005;
Winter et al., 1995). A modified P–M equation is also used to com-
pare bare soil evaporation and transpiration from deep- and shal-
low-rooted vegetation as described below. Average air temperature,
soil moisture, and mean sea level pressure, relative humidity and
average wind speed obtained from MESonet were used to cal-
culate P–M reference ET rates.

Bare soil evaporation was assumed to remove moisture from
the unsaturated zone and it is applied only to the portion of soil
with shallow-rooted vegetation or bare soil cover.

\[ e_{bs}(t) = (1 - M)ke_{bs}e_{mod}(t) \]  
(8)

where $e_{bs}$ is the bare soil evaporation in cm/day, $M$ is the por-
tion of land covered by deep-rooted vegetation or phreatophytes, $ke_{bs}$
is a bare-soil evaporation scaling factor and $e_{mod, PM}$ is the actual eva-
poration taking the availability of soil moisture into account. $e_{mod, PM}$
was calculated using the P–M equation as a reference ET rate and
modified as described in Jia et al. (2001) to account for soil moisture
availability. The method is based on the energy balance of the soil
surface, aerodynamic diffusion equations of latent and sensible heat
fluxes and the wetness function concept (Lee and Pielke, 1991; Noil-
han and Planton, 1988). The equation as derived by Jia et al. (2001)
is given below.

\[ e_{mod, PM}(t) = \begin{cases} 
0 & \theta(t) \leq \theta_m \\
\frac{\Delta h_{\text{mod}}(t)}{G_{\text{mod}}(t)} & \theta_m < \theta(t) < \theta_c \\
\frac{1}{D_{\text{mod}}(t)}(\theta_c - \theta(t)) & \theta(t) \geq \theta_c
\end{cases} \]  
(9)

We note that $e_{mod, PM}(t) = e_{mod, PM}(t)$ if $\beta(t) = 1$. In the above, $\theta$
is the soil moisture content, $\theta_c$ is the field capacity of the top soil
layer, $\theta_m$ is the soil moisture content corresponding to the mono-
molecular suction, $H_{\text{sat}}$ is the net radiation (MJ/m$^2$ day), $G$ is the
soil heat flux (MJ/m$^2$ day) (assumed to be negligible since it has
small values from day to day because heat stored in the day is
lost at night), $\rho_{sat}$ is the air density (kg/m$^3$), $C_p$ is the specific
heat of moist air at constant pressure, $e^*_0$ is the saturation vapor
pressure of air at height $z$ (kPa), $e_c$ is the actual vapor pressure
of air at height $z$ (kPa), $r_c$ is the canopy surface resistances (for
bare soil there is no canopy resistance and $r_c = 0$ day/m), $\gamma$ is
the aerodynamic resistance to sensible heat and vapor transfer
(day/m), $z$ is the plant height ($\text{m}$), $\Delta$ is the slope of the satu-
rator suction pressure–temperature curve (kPa/°C), $\gamma$ is the psy-
chometric constant (kPa/°C), $\beta$ is the wetness function based on
the work of Jia et al. (2001), and $e_{mod}$ is the reference evaporation
rate calculated with the P–M equation (kg m$^{-2}$ d$^{-1}$) (Eq. (9) was
divided by the density of water to obtain units cm/day). Following Jia et al. (2001), the wetness function is estimated with
Eq. (10).

\[ \beta(t) = \frac{1}{4} \left( 1 - \cos(\pi(\theta(t) - \theta_m)/(\theta_c - \theta_m)) \right)^2 \]  
(10)

Eq. (9) indicates that evaporation cannot take place when soil mois-
ture $\theta(t)$ is less than the monomolecular suction $\theta_m$. When soil
moisture falls between $\theta_m$ and the field capacity $\theta_c$, evaporation
is given by the modified formula based on the wetness concept. Final-
ly, the P–M equation is used to compute evaporation when soil
moisture exceeds the field capacity in which case the land surface
effectively behaves as a free water surface. For bare soil there is
no canopy resistance thus $r_c$ is set to zero. From Eqs. (8) and (9) it
is observed that the P–M equation with $r_c = 0$ gives an upper limit
for evaporation rates therefore the P–M equation is expected to
be a reference evaporation rate that is modified according to mois-
ture conditions in the soil.

Transpiration from vegetation in the unsaturated zone ($e_{cv,unsat}$)
used the P–M ET ($e_{PM}$) as a reference rate and it was applied only
to the portion of land ($M$) covered by phreatophyte vegetation that
uses both soil water and groundwater.

\[ e_{cv, unsat} = k_{vg} \rho(t)(1 - R(t))e_{PM}(t) \]  
(11)

where $\rho$ denotes the plant water stress function (Muneepeerakul
et al., 2008), $k_{vg}$ is a transpiration efficiency factor and $R$ denotes
the proportion of roots above the water table which is defined as an
exponential function with a specific mean root depth (MRD) (Muneepeerakul et al., 2008):

\[ R(t) = \exp \left[ -\frac{1}{\text{MRD}} \frac{y(t)}{\rho} \right] \]  
(12)

Plant water stress ($\rho$) is given by:

\[ \rho(\theta) = \begin{cases} 
1 & \theta(t) > \theta^* \\
\frac{\theta(t)}{\theta_u - \theta_w} & \theta_u < \theta(t) < \theta^* \\
0 & \theta(t) < \theta_u
\end{cases} \]  
(13)

where $\theta_u$ is the soil moisture at which stomatal closure is fully
complete, and $\theta^*$ is the soil moisture at which stomatal closure begins
($\theta^* = \theta_h$). $\theta_u$ and $\theta^*$ are defined as water saturation fractions of
the pore space.

After groundwater has been recharged, water is lost from the
saturated zone via transpiration from phreatophyte vegetation that
uses water from the saturated and unsaturated zones. Transpira-
tion from vegetation in the saturated zone is applied only to the
portion of land covered by deep-rooted vegetation ($M$) or
phreatophytes:

\[ e_{sv, sat} = M_{v, sat} k_{g, sat} R(t) e_{PM} \]  
(14)

where $e_{sv, sat}$ is the transpiration from the saturated zone (cm/day),
$k_{g, sat}$ is the plant transpiration efficiency in the saturated zone.

4.2.1.2. Groundwater flow components and water table fluctua-
tions. Fluctuations of water table depth were modeled based on
the water balance in the riparian catchment as follows.

The water table depth ($y$) defined in the model is the distance
from the ground surface to the groundwater level. Therefore $y$
decreases with recharge ($R_e$), and increases with transpiration
($e_{sv, sat}$), the local groundwater flow component ($q_{local}$), and the
regional groundwater flow ($Q_{reg}$):

\[ y(t + \Delta t) = y(t) - \frac{R_e(t)}{\phi(1 - \theta_c)} + \left[ e_{sv, sat}(t) + q_{local}(t) + Q_{reg}(t) \right] \Delta t \]  
(15)

where $y$ is the water table depth. Local and regional groundwater
flow components were modeled to flow in response to the differ-
ences in head between the wetland and local groundwater levels
(Krasnostein and Oldham, 2004).
Local subsurface flow occurs when a difference of head between the riparian catchment and the wetland exists. A response time, $t_{\text{Rlocal}}$, for the subsurface flow component was calculated similar to the approach followed by Jothityangkoon et al. (2001) and Krasnostein and Oldham (2004), which is based on Darcy’s flow velocity. In this study we used two buckets, one for representation of the riparian system and the other for the wetland system. Both buckets contain a regional groundwater flow that affects the baseline water level over the extent of the larger flow system, which is connected to the main river. The boundary of the regional system is defined by the extension of the aquifer. The connection between the riparian bucket and the wetland bucket is the groundwater flow from the riparian area to the wetland. Since they are separate bucket systems, Darcy’s Law cannot be applied directly, instead we needed to find the response time for the excess of water from one bucket (riparian area) to reach the other bucket (the wetland area). Then we used Darcy’s velocity (Eq. (16)) and characteristic length of the riparian area to calculate a response time and the excess of water is represented as the difference of water levels between the riparian area and the wetland area.

$$v_{\text{Rlocal}}(t) = \frac{K_{\text{GWlocal}} g}{\rho} i_{\text{Rlocal}}(t) \quad (16)$$

where $K_{\text{GWlocal}}$ is the characteristic hydraulic conductivity used to calculate subsurface local groundwater flow (cm/day), and $i_{\text{Rlocal}}$ is the time-dependent hydraulic gradient

$$i_{\text{Rlocal}}(t) = \frac{\Delta h(t)}{L_{\text{GWlocal}}} \quad (17)$$

where $L_{\text{GWlocal}}$ is given in cm and is the characteristic length of the idealized representation of the system (distance between the riparian area and the wetland) that is used to calculate local groundwater flow. $\Delta h$ is the time varying head difference between wetland and local groundwater levels in cm in the riparian area. Accordingly, the response time, $t_{\text{Rlocal}}$ is given by:

$$t_{\text{Rlocal}}(t) = \frac{L_{\text{GWlocal}}}{v_{\text{Rlocal}}(t)} \quad (18)$$

The local subsurface flow component is then calculated as the head difference divided by the response time:

$$q_{\text{local}}(t) = \frac{\Delta h(t)}{t_{\text{Rlocal}}(t)} \quad (19)$$

The subsurface flow that feeds the regional groundwater flow depends on the physical properties of the entire system: regional hydraulic gradient ($i_{\text{reg}}$), which could be obtained from water levels in the regional area reported in former studies (Scholl and Christenson, 1998), regional hydraulic conductivity ($K_{\text{GWreg}}$) and porosity ($\rho$):

$$Q_{\text{reg}}(t) = K_{\text{GWreg}} i_{\text{reg}}(t) \quad (20)$$

where the subscript “reg” represents the regional hydraulic properties. The model makes a distinction between the local and regional flow components by allowing the hydraulic conductivity and the hydraulic gradient to be different for the two processes, therefore the regional hydraulic gradient will be a function of a regional length scale $L_{\text{GWreg}}$:

$$i_{\text{reg}}(t) = \frac{\Delta h(t)}{L_{\text{GWreg}}} \quad (21)$$

where $L_{\text{GWreg}}$ is given in cm and is the characteristic length used to calculate regional groundwater flow and $\Delta h$ is the time varying head difference between wetland and local groundwater levels in cm.

### 4.2.2. Water balance in the wetland area

The wetland is assumed to receive water from precipitation ($p$) and groundwater flow from the riparian catchment ($q_{\text{local}}$) and lose water through ET ($e_{\text{WL}}$), regional groundwater flow $Q_{\text{reg}}(t)$ and local subsurface outflow to the aquifer ($q_{\text{out}}$):

$$y_{\text{WL}}(t + \Delta t) = y_{\text{WL}}(t) + \left[ p(t) + R(t) - e_{\text{WL}}(t) - Q_{\text{reg}}(t) + q_{\text{local}}(t) \frac{A_1}{A_2} - q_{\text{out}}(t) \right] \Delta t \quad (22)$$

where $y_{\text{WL}}$ is the simulated water level depth in the wetland area (cm), $A_1$ and $A_2$ denote the areas of the riparian zone and wetland respectively. The change in water level of the wetland is dependent on the area-volume relation of the wetland. This relation was based on data from detailed bathymetry surveys for the wetland and is represented as polynomial fits to the depth-area and depth-volume data. The ratio between $A_1$ and $A_2$ represents changes in storage across the system due to changes in bucket volumes (Krasnostein and Oldham, 2004). Flow-through wetlands in dunal sands have been found to intercept groundwater from the top 6–10 m of an aquifer even though the wetlands have been shallow (Smith and Townley, 2002). The unconfined alluvial aquifer reported in this paper is 10–15 m thick, therefore it is expected that the regional groundwater flow ($Q_{\text{reg}}$) in the area will uniformly drop the water table elevation as the water moves to the Canadian River. The regional groundwater flow ($Q_{\text{reg}}$) will affect the water level in the wetland as it does in the riparian area. Local subsurface outflow from the wetland to the aquifer ($q_{\text{out}}$) was calculated as a subsurface flow triggered by the difference in hydraulic heads of the system. A response time taken to release the excess water from the wetland through the soil bank, $t_{\text{out}}$, for the local subsurface outflow component was calculated based on the head gradient $i_{\text{out}}$:

$$i_{\text{out}}(t) = \frac{\Delta h(t)}{L_{\text{GWout}}} \quad (23)$$

where $L_{\text{GWout}}$ is given in cm and is the characteristic length used to calculate local groundwater outflow from the wetland to the aquifer and $\Delta h$ is the time varying head difference between wetland and local groundwater levels in cm. Accordingly the response time, $t_{\text{out}}$ is given by:

$$t_{\text{out}}(t) = \frac{L_{\text{GWout}}}{v_{\text{out}}(t)} \quad (24)$$

$$v_{\text{out}}(t) = \frac{K_{\text{GWout}} g}{\rho} i_{\text{out}}(t) \quad (25)$$

where $K_{\text{GWout}}$ is the characteristic hydraulic conductivity used to calculate subsurface local outflow from the wetland to the aquifer (cm/day). The local subsurface outflow component was then calculated as the difference in head divided by the response time:

$$q_{\text{out}}(t) = \begin{cases} \frac{\Delta h(t)}{t_{\text{out}}(t)} & \text{for } y_{\text{WL}}(t) > 0 \\ 0 & \text{for } y_{\text{WL}}(t) \leq 0 \end{cases} \quad (26)$$

ET losses ($e_{\text{WL}}$) in the wetland were calculated using ($e_{\text{WL}}$) as a reference rate, and a plant efficiency factor ($k_{\text{WL}}$) (Farmer et al., 2003):

$$e_{\text{WL}} = k_{\text{WL}} e_{\text{WL}} \quad (27)$$

The governing equations of the model were solved using the forward Euler scheme (Press et al., 2007) with a daily time step because meteorological data for long periods of time were collected daily. Model parameters were estimated by minimizing the following objective function $F$:

$$F = F_{\text{GW}} + F_{\text{WE}} \quad (28)$$

$$F_{\text{GW}} = \sum_{t=1}^{\text{year}} \left( \frac{\dot{y}_{\text{obs}} - \dot{y}_{\text{WL}}}{\dot{y}_{\text{obs}}} \right)^2$$

$$F_{\text{WE}} = \sum_{t=1}^{\text{year}} \left( \frac{\dot{y}_{\text{WL}} - \dot{y}_{\text{WL}}}{\dot{y}_{\text{WL}}} \right)^2$$
where \( y \) and \( y_{WL} \) are the simulated water table depth and water level depth in the wetland area respectively (cm), \( y_{obs} \) and \( y_{WL, obs} \) are the observed water table depth and water level depth in the wetland area respectively (cm), \( FGW \) is the error between simulated and observed water table depths for groundwater and \( FWL \) is the error between simulated and observed water levels in the wetland.

After identifying the important parameters using sensitivity analysis, model parameters were estimated using a direct search method for solving optimization problems implemented in MATLAB (Version R2009A, The Mathworks Inc., Natick, MA). A list of parameters estimated and the best set of parameter values is listed in Table 1.

5. Results and discussion

5.1. Water balance model

Simulated water table depths in the riparian area and water levels in the wetland were compared with observed data to evaluate the model (Fig. 4). The comparison shows good overall agreement with observed data. Model predictions of water levels in the wetland for the extended period December 2006–January 2009 are also included in Fig. 4 (a vertical line separates the two different simulation periods). A reasonable agreement was obtained using the parameters estimated for the period May 5 1997–December 4 2006.

Historically, it was observed that water levels in the wetland and water table depths varied seasonally decreasing during summer, reaching minimum levels during fall, followed by an increase during winter to maximum levels in spring (Báez-Cazull et al., 2007). The model was able to describe this seasonal behavior in the wetland and the riparian area of the site (Fig. 4). Differences in ET rates between the unsaturated zone and the saturated zone as well as between the riparian area and the wetland were considered in the model; this aspect of the model could be useful in evaluating how different vegetation types influence the water balance in the wetland in different climatic settings.

Results showed that an important event occurred between November 16, 2005 and March 16, 2006. Simulations for both variables (water table depth and water level in the wetland) predicted that the system would be drying out while measured water levels tended to increase during the period (Fig. 4). Measured precipitation recorded low rates for that period with a maximum of 0.7 cm/day and 0.02 cm/day on average. The lack of precipitation correlates well with the drying out prediction of the model since the model assumes that precipitation is the only source of water to the system. Extreme climatologic conditions were reported during fall 2005 and winter 2005–2006 around the area. Fall 2005 was drier than normal and was characterized by lack of precipitation (McManus, 2006b). Winter 2005–2006 was dry and warm as well, however some extreme cold fronts generated snow in some areas followed by warm temperatures, low humidity, strong winds and dangerous wildfire conditions (McManus, 2006a). Oklahoma City, OK (north of Norman, OK) broke its record for daily snowfall on December 20, 2005 with 4.1 cm (1.6 in.) (McManus, 2006a), followed by temperature rise upwards of 70 °F in some portions of the state. Surface runoff, subsurface flow or a combination of both due to snow melt periods are potential sources of water that the model did not account for.

5.1.1. Model parameters

Results of soil moisture in the unsaturated zone as well as infiltration that remained in the unsaturated zone (\( k_{US} \)), bare soil evaporation (\( e_{BS} \)) and transpiration from vegetation in the unsaturated zone (\( e_{veg} \)) are depicted in Fig. 5. Field measurements of soil moisture were not conducted at the site, thus Fig. 5 contains only the measured values.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Best estimated value</th>
</tr>
</thead>
<tbody>
<tr>
<td>( k_{US} ) (cm d(^{-1}))</td>
<td>233.33</td>
</tr>
<tr>
<td>( k_{LS} ) (cm)</td>
<td>4875.70</td>
</tr>
<tr>
<td>( k_{SR} ) (cm d(^{-1}))</td>
<td>1.0</td>
</tr>
<tr>
<td>( y_{WL} ) (cm)</td>
<td>0.61</td>
</tr>
<tr>
<td>( L_{C_{LS}} ) (cm)</td>
<td>3332.00</td>
</tr>
<tr>
<td>( M_{RD} ) (cm)</td>
<td>253.70</td>
</tr>
<tr>
<td>( k_{US} ) (cm)</td>
<td>2.63</td>
</tr>
<tr>
<td>( k_{LS} )</td>
<td>1.36</td>
</tr>
<tr>
<td>( L_{C_{SR}} ) (km)</td>
<td>30.00</td>
</tr>
<tr>
<td>( M ) (km)</td>
<td>0.74</td>
</tr>
<tr>
<td>( k_{SR} ) (cm d(^{-1}))</td>
<td>176.10</td>
</tr>
<tr>
<td>( \varphi )</td>
<td>0.31</td>
</tr>
<tr>
<td>( k_{veg} )</td>
<td>0.90</td>
</tr>
<tr>
<td>( k_{BS} )</td>
<td>0.27</td>
</tr>
<tr>
<td>( h ) (cm)</td>
<td>0.60</td>
</tr>
<tr>
<td>( h_{w} )</td>
<td>0.59</td>
</tr>
</tbody>
</table>

Table 1
Summary of best set of estimated model parameters.
simulated results. Since the model does not allow for soil moisture to raise above field capacity the maximum soil water moisture that will remain in the soil after infiltration is the value of field capacity. The highest value of soil moisture should be equal to the field capacity ($h_{fc}$) and the lowest value should be between the soil moisture value at which stomatal closure begins ($\theta'$) and moisture content corresponding to monomolecular suction ($\theta_m$). In the model, $\theta'$ and $\theta_m$ values are lower limits for transpiration from vegetation and bare soil evaporation respectively, i.e., below these thresholds no evaporation or transpiration takes place. The best set of our estimated parameters provided a narrow range for $h_{fc}$ and $h_{C3}$ as well as $h_{fc}$ and $h_m$ thus resulting in a narrow range of soil moisture values in the unsaturated zone of the riparian area (Fig. 5a). A narrow range of soil moisture values appears reasonable since in the riparian catchment the soil is mostly composed of sand with some silt, clay and lenticular beds of gravel (Scholl et al., 2005).

Fig. 5b depicts the amount of precipitation infiltrated that contributes to soil moisture, which depends on the capacity of the soil to hold water. Water storage capacity of the soil in the unsaturated zone is a function that depends on the porosity, field capacity and time-dependent soil moisture. Estimated porosity value (0.31) is around the lower limit of measured values (0.34–0.44). Field capacity was not measured at the site, however the estimated value of 0.61 appears reasonable since it is similar to the value used for loamy sand (0.52) reported in Muneepreeakul et al. (2008).

The water balance in the riparian area is composed of recharge, transpiration from deep-rooted plants in the saturated zone, a subsurface flow that contributes to the regional groundwater flow and local groundwater flow component. Each component of the water balance in the riparian area is depicted in Fig. 6. Transpiration from the saturated zone is on average 3.7 times larger than transpiration from the unsaturated zone. This can be explained based on the fact that deep-rooted phreatophytes that inhabit this area (willow, cottonwood, Eastern red cedar, and salt cedar) are acclimated to uptake water (via tap roots) from the saturated zone, which seems to be a more dependable source of water than soil moisture in the sandy soil of the study site.

Each component of the water balance of the wetland area is depicted in Fig. 7. In the bucket model approach local groundwater flow is multiplied by the ratio ($R_{A1}$) between the riparian area ($A_1$) and the wetland area ($A_2$) since change in storage across the system will vary according to relative changes in bucket volume (Kranzstein and Oldham, 2004). In Fig. 7c results from local groundwater flow are presented multiplied by the ratio $R_{A1}$. It can be observed that high peaks in $q_{local}(t)/[A_1/A_2]$ correspond to the combination of high peaks in $q_{local}$ with low stage values in the wetland. Therefore, the contribution of local groundwater flow to

![Simulated results for soil moisture in the riparian area.](image-url)

*a Simulated results for soil moisture in the riparian area. Components of soil moisture in the water balance model are: (b) change in water storage in the unsaturated zone, that is infiltration ($I_{US}$) when $I_{US} < WSC$ or water storage capacity ($WSC$) when $I_{US} > WSC$), (c) bare soil evaporation and (d) transpiration in the unsaturated zone.*
the wetland depends on the volume difference between the riparian catchment that funnels water to a smaller wetland area.

Subsurface flow components ($Q_{reg}$, $q_{local}$, $q_{out}$) were modeled based on a time response that depended on characteristic length scales. These length scales can be linked to the length of the catchment in the groundwater flow direction. Values from parameter estimation of local ($L_{GWlocal}$) and outflow ($L_{GWout}$) characteristic lengths were 33.32 m and 48.75 m respectively. These values are close to the approximate length of the catchment. Regional characteristic length ($L_{GWreg}$) on the other hand was noticeably larger, which corroborates the conceptual framework that the aquifer–wetland system is part of a larger flow system connected to the main river. Hydraulic conductivity values measured at the site are highly variable. As mentioned earlier in the materials and methods section, a study of spatial variation of hydraulic conductivity using slug tests (Scholl and Christenson, 1998) reported a range from $8.4 \times 10^{-7}$ to $2.8 \times 10^{-8}$ m/s with a median value of $6.6 \times 10^{-5}$ m/s. Parameter estimation provided values of $2.7 \times 10^{-5}$ m/s, $1.2 \times 10^{-7}$ m/s, and $2.1 \times 10^{-8}$ m/s for local, regional and outflow hydraulic conductivity respectively. Local and outflow hydraulic conductivities obtained from parameter estimation were within the range of reported conductivities.

5.1.2. Vegetation parameters

Bare soil evaporation in the unsaturated zone of the riparian area is depicted in Fig. 5c. Bare soil evaporation was obtained using P–M evaporation (with $r_c = 0$) as a reference rate. The reference rate was modified by a scaling factor ($k_{BES}$) that accounts for higher evaporation rates in the case where shallow-rooted vegetation covers the bare soil area (Farmer et al., 2003). A value of 1.36 was obtained for $k_{BES}$ from parameter estimation suggesting that the riparian area is covered by grass. This appears to be reasonable since shallow-rooted vegetation of the riparian area is mainly composed of western ragweed, Bermuda grass, Johnson grass, and bundle flower.

Bare soil evaporation and transpiration rates in both the unsaturated and saturated zones of the riparian area depend on the percentage of soil covered by deep-rooted vegetation ($M$). A $M$ value of 0.73 was obtained from parameter estimation. Expected deep-rooted vegetation coverage of 0.44 was derived from the National Land Cover Data 2001 (NLCD2001) (available at http://www.epa.gov/mrlc/nlcd-2001.html). The discrepancy may be due to the fact that the land cover type used to obtain the expected value of $M$ was deciduous forest; however for this model additional land cover types in the vegetation coverage calculation might be necessary to properly reflect the heterogeneous vegetation cover. Transpiration
in the unsaturated zone was obtained using P–M as a reference rate modified by a plant transpiration efficiency factor \( k_{\text{veg}} \) for deep-rooted vegetation cover (Jothityangkoon et al., 2001) in the unsaturated zone. It is recognized that soil water uptake is a function of total water potential, which include gravitational, matrix, and osmotic potentials and not only a function of the root. Consideration of all inhibition sources requires a detailed modeling of vegetation dynamics, which is beyond the scope of the present paper. A value of 0.3 for plant transpiration efficiency was obtained in the model and it may suggest that deep-rooted vegetation in the riparian area is not acclimated to use soil moisture from the unsaturated zone. Deep-rooted vegetation in the riparian area is composed of at least three species of facultative phreatophytes that in the presence of shallow groundwater are more likely to use groundwater and significantly lower the water table through high rates of transpiration from the saturated zone (Burgess, 2004).

Transpiration in the unsaturated and saturated zones of the riparian area (Fig. 6c) was modeled as a function that depends on the proportion of land covered by deep-rooted vegetation \( M \), proportion of roots above the water table \( R \), and a plant transpiration efficiency factor \( k_{\text{veg,sat}} \). The transpiration factor can be regarded as an anoxic coefficient that reflects adverse effects such as reduced photosynthesis activity from increased stomatal closure, toxicity of reduced minerals, among others (Muneepeerakul et al., 2008). The value of 0.9 obtained from parameter estimation may reflect that phreatophytes in the site are acclimated for anoxic conditions since the water table has been relatively shallow for over 9 years.
5.1.3. Recharge compared to independent measurements

To understand the importance of net recharge in the water balance we expressed recharge (Re) as a percentage of precipitation \(\text{Re}(t)/p(t)\) ignoring delay in the unsaturated zone (a reasonable assumption for long-term estimates) and calculated 10-year and annual averages for the period of study. Over the 10 year period of study we found that recharge to shallow groundwater is on average around 36% of rainfall at this site. The percentage of rainfall that recharges the saturated zone varies seasonally (Fig. 8a) – during winter and fall approximately 43% and 44% of rainfall eventually becomes recharge. During spring the percentage of precipitation that recharges groundwater is at the lowest with 27% followed by summer with 29%. This seasonal trend can be related to growing seasons, where rainfall is readily transpired by vegetation during spring and summer leaving relatively small amounts of water free for recharge. Transpiration rates decline during fall and winter allowing larger percentages of rain to become recharge. Fig. 8b shows the seasonal net recharge as a percentage of precipitation of the riparian area from 1997 to 2009. In this plot it is possible to note the important annual signatures. The year 2000 is characterized by a high percentage of rain recharging the groundwater with winter and fall having the largest percentages from the 10 years of study. The year 2005 is characterized by a low percentage of rain becoming recharge, and with the distinctive feature that contrary to the 10-year average trend, summer has a larger percentage of recharge compared to winter. A study on recharge processes using isotopes at this site (Scholl et al., 2005) reported values of 16%, 22%, and 64% of recharge as

Fig. 8. Recharge as a percentage of precipitation in the riparian area: (a) 10 year average of summer, fall, winter and spring seasons, and (b) seasonal averages for 1997–2009. Red lines plotted in the period highlighted (fall 1998 to spring 2000) correspond to estimates of recharge from Scholl et al. (2005). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
a percentage of rainfall (depending on the method used) for the period from October 1998 to May 2000. Our values are similar to the values reported in Scholl et al. (2005), an average of 34% is observed from fall 1998 to spring 2000, a maximum of 51% was observed for winter 1998 and a minimum of 27% for spring of 1999.

5.2. Groundwater inflow and outflow to the wetland and seepage measurements

5.2.1. Groundwater flow compared to independent measurements

Subsurface flow components are seasonal, and readily respond to rainfall events (Fig. 7e and f). Regional groundwater flow ($Q_{\text{reg}}$) contribution is relatively small compared to local subsurface flow ($q_{\text{local}}$). This is reasonable since the conceptual model considers that the riparian area and wetland area are part of a larger flow system connected to the main river.

Seepage meters in the wetland measured the net groundwater inflow or outflow. Measurements were conducted in approximate intervals of 7 days in 2005 using a total of 27 seepage meters installed in the wetland bed. Individual meter measurements were used to obtain an average for each interval. Model subsurface flow can be divided into regional and local outflow from the wetland and inflow to the wetland. Modeled net groundwater inflow to the wetland was obtained by subtracting the outflow $Q_{\text{reg}}(t) + q_{\text{out}}(t)$ from the inflow $q_{\text{local}}(t)/A_1/A_2(t)$. Comparison between measured (seepage meters) and modeled net groundwater inflow to the wetland for the year 2005 is depicted in Fig. 9 which shows good agreement between modeled and observed values of net groundwater inflow to the wetland.

![Comparison between simulated groundwater inflow rates and observations based on seepage measurements](image1)

![Comparison between simulated groundwater inflow rates and observations based on seepage measurements](image2)

Fig. 9. Comparison between simulated groundwater inflow rates and observations based on seepage measurements (each observed value represents the average based on measurements from 27 seepage meters installed at the wetland bed).

Fig. 10. (a) Isotope composition of rainfall, wetland, and well samples. $^{18}\text{O}$ versus $^2\text{H}$ graph values for rainfall, wetland sampled close to transect 148 (Wetland 148), groundwater sampled from well 148C (well 148C), wetland sampled close to transect 153 (Wetland 153) and groundwater sampled from well 153C (well 153C). (b) Comparison between simulated and observed (isotope mass balance approach) groundwater inflow and outflow rates for transect 148.
Modeled groundwater inflow to and outflow from the wetland were compared to groundwater flows calculated with the unsteady state isotope mass balance approach explained in Appendix A. We used isotope concentrations of groundwater samples from well C of transect 148 (Fig. 1), wetland water samples from locations close to the transect and rainfall samples collected approximately every month from March 20, 2007 to September 15, 2008. To evaluate isotope characteristics of different water samples a plot of \( ^{18}\text{O} \) versus \( ^{2}\text{H} \) is shown in Fig. 10a. The local meteoric water line (LMWL) shown in Fig. 10a was obtained from a previous study at the site (Scholl et al., 2006) that used 3 years of biweekly rainfall samples (from May 1996 to May 2000). In general rainfall samples lie along the LMWL. The evaporative line (EL) in Fig. 10a represents the evaporation trend and is a regression line based on isotopic analyses of 30 wetland samples. It can be observed that \( ^{18}\text{O} \) in water samples from the wetland are generally offset below the LMWL due to evaporation. In samples from the wells (Well 148C) \( ^{2}\text{H} \) is generally offset above the LMWL suggesting that these samples contain leachate and the enrichment is probably due to methanogenesis occurring in the aquifer (Hackley et al., 1996; Scholl et al., 2006). Thus three different types of water can be distinguished, namely: rainfall, groundwater and wetland. Daily rates from modeled groundwater inflow (q_{inflow}(t)/A_1(t)) and outflow (q_{out}(t)) to the wetland are compared with the corresponding isotope measurements in Fig. 10b. While some differences between observed and simulated inflows and outflows were noted around June 7, 2007, the agreement is generally modest and similar trends were noted in both datasets.

5.2.2. Evaporation rates compared to independent measurements

In the model, bare soil evaporation and transpiration in the unsaturated zone depend on the amount of moisture available. In general, transpiration rates are four times smaller than bare soil evaporation (Fig. 5c and d). According to the conceptual model, where water in the soil is no longer available for evaporation and transpiration to occur when soil moisture is below the low limits, this result suggests that there is not enough soil moisture in the sands available for plants to uptake. Conditions of soil moisture at low capacity limits where soil evaporation and transpiration are low or null arise during periods of no rainfall. A soil moisture deficit will evolve since bare soil evaporation and transpiration remove water up to the low limits (\( \theta^*, \theta_m \)). In the model it is assumed that when soil moisture is below the low limits, water in the soil is no longer available for evaporation and transpiration to occur.

Fig. 10a and b depict comparisons between modeled ET and measured evaporation rates in the wetland and the riparian area. Evaporation rates measured with the Class A pan placed on land (land-pan measured) were compared to P–M evaporation rate with a zero canopy resistance (\( r_c = 0 \)) for bare soil (Farmer et al., 2003). Comparison between land-pan evaporation rates and P–M evaporation rates shows that both rates are in good agreement. These comparisons support the well-known result that potential ET for short vegetation is very similar to free water evaporation (Brutsaert, 1982).

ET rates measured with a float pan in the wetland were compared with P–M ET rates since these values were used as reference ET in the wetland water balance (Fig. 11). It can be observed that P–M ET rates are very similar to the measured float pan values. Modeled ET rates (Fig. 7) for the entire wetland area are around 3.7 times larger than float-pan measurements and 2.64 \( (k_{wl}) \) times larger than P–M ET. The difference comes from the fact that evaporation rates measured with a floating pan submerged in the wetland simulate physical conditions that control evaporation in the wetland free-water surface (Masoner et al., 2008) while ET rates modeled in the wetland area implicitly involve open water evaporation and transpiration from wetland vegetation such as cattail. The \( k_{wl} \) value of 2.64 in the model can be related to ET crop coefficients that account for specific physiology of plant species (Allen et al., 1998). In a study of crop coefficients for wetland vegetation Towler et al. (2004) found coefficients of around 3.29–5.91 for broad leaf cattail and hardstem bulrush, therefore our estimated value of 2.64 appears to be reasonable for the study site.

While state variables in some process-based bucket models described observed data reasonably well in the past (Krasnostein and Oldham, 2004; Farmer et al., 2003), to the best of our knowledge none of them attempted to evaluate modeled fluxes using independent estimates based on observations in a wetland system. Similar to Krasnostein and Oldham (2004) model simulations were used to identify input and output fluxes to the wetland water budget, such as ET. In the present work, we were able to quantify evaporative fluxes as well as recharge using additional data from evaporation pans.

Fig. 11. Simulated results for evapotranspiration rates (a) in the wetland area and (b) the riparian area. Penman–Monteith evapotranspiration rates were used as reference evapotranspiration rates in the model.
ET is an important component in the water balance of wetlands. In Kranostein and Oldham (2004) bare soil evaporation is differentiated from ET based on the percentage of deep-rooted vegetation in the site. In their study evaporation and ET rates were calculated based on a response time of the catchment with respect to evaporation. The response time depended on the relative water storage at the given time step. Our model differs from that of Kranostein and Oldham (2004) in that ET rates were modeled to reflect unique characteristics of bare soil evaporation and transpiration in the riparian area as well as ET in the wetland area. Our model presents an advance to that of Kranostein and Oldham (2004) and Munepeerkul et al. (2008) in that it could be used to evaluate the effect of changes in vegetation communities on the water storage of the wetland, since it differentiates between catchment and wetland ET fluxes. This aspect of the modeling is important since wetland vegetation is expected to be significantly different from vegetation in the catchment.

6. Conclusions

Our results indicate that subsurface flow components at the Norman landfill site are seasonal and readily respond to rainfall events. Regional groundwater flow contribution to the wetland is relatively small compared to local subsurface flow. On average, recharge at the site over a 10 year period was found to be approximately 40% of rainfall. Using the simple bucket-type modeling framework, we were able to explore linkages between the hydrologic regime of the wetland and its relationship to climate, soil and vegetation dynamics. Rather than depending on measurements for each component, the model mainly relies on climatic data. Model parameters were constrained using field measurements at the site. The model is conceptually simple, yet it was able to adequately describe fundamental hydrological processes. A recharge study at the site (Scholl et al., 2005) suggested that recharge to the riparian area is around 16–64% of rain. Results from the water balance model presented here corroborate this suggestion.

In this study important hydrological processes were quantified such as: the inflow, outflow, recharge and ET rates. The knowledge of these processes is important to further quantify nutrient, carbon, sediment, and/or contaminant cycling in the wetland. Results from the model can be used to determine the rate and volume of groundwater entering the wetland as well as the origin of the water. Model results of the response of an aquifer–wetland system to climate, soil and vegetation were useful to understand the intensity and duration of groundwater recharge, the groundwater inflow and outflow rates and to and from the wetland and differences in ET rates between the riparian area and the wetland. This long-term information will be useful in evaluating the effects of climate change on dominant hydrological processes as well as the capacity of the wetland to degrade persistent contaminants. While the present paper outlines a method of integrating long-term data with simple models, we note that the data collected may not include the full range of variability experienced under climate change and some of the processes could potentially change leading to decreased model skill. The approach, however, is expected to be useful as it is relatively easy to examine alternative conceptual models within the bucket model framework and see how different fluxes change in time.

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Appendix A. Groundwater inflow calculations using isotopes

Many wetlands cannot be considered to be at hydrologic or isotopic steady state since they receive pulses of water. An approach is needed where the steady state assumption can be relaxed. The general water mass balance relationship for a wetland is:

\[
\frac{dV}{dt} = Q_{GW} + Q_P - Q_e - Q_{OGW}
\]

where \(V\) is the volume of the wetland, \(t\) is the time, \(Q_{GW}\) is the groundwater inflow to the wetland, \(Q_P\) is the precipitation inflow to the wetland, \(Q_e\) is the evaporation outflow from the wetland and \(Q_{OGW}\) is the groundwater outflow from the wetland. Similarly the isotope mass balance for a wetland is:

\[
\frac{d(V C_e)}{dt} = Q_{GW} C_{GW} + Q_P C_P - Q_e C_e - Q_{OGW} C_{GW}
\]  

(A2)

where \(C_e\), \(C_{GW}\), \(C_P\), and \(C_{OGW}\) are the isotope concentration of the water in the wetland, groundwater inflow to the wetland, precipitation inflow to the wetland and \(Q_{GW}\) is the groundwater outflow from the wetland. Similarly the isotope mass balance for a wetland is:

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\frac{d(V C_e)}{dt} = Q_{GW} C_{GW} + Q_P C_P - Q_e C_e - Q_{OGW} C_{GW}
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\[
\frac{d(V C_e)}{dt} = Q_{GW} C_{GW} + Q_P C_P - Q_e C_e - Q_{OGW} C_{GW}
\]  

(A2)
position of atmospheric water vapor given by (Yehdegho et al., 1997):

$$\delta_A = \delta_A \left( \frac{1}{A} \right) - \left( 1 - \frac{1}{A} \right)$$

(A5)

where $A$ is obtained by (Majoube, 1971):

$$10^3 \ln x = 2.400 \times 10^7 - 6.455 \times 10^5 - 1.167$$

Finally, to obtain the concentration of evaporative flux in ratio notation from Eq. (A3):

Assuming well-mixed conditions in the wetland, groundwater outflow concentration is the same as the concentration in the wetland $C_{GW} = C_i$. Solving Eq. (A1) for $Q_{GW}$, substituting into Eq. (A2) and solving Eq. (A2) for $Q_{GW}$ we obtain:

$$Q_{GW} = \frac{Q_i (C_i - C_f) + Q_2 (C_i - C_f) - V \delta_A (C_f - C_{GW})}{\delta_A}$$

(A9)

Volumetric precipitation flow rate ($Q_p$) into the wetland depends on the precipitation rate ($P$) and the surface area of the wetland $A(H) = 12128.89 \text{m}^2$

$$Q_p = PA(H)$$

(A10)

where $H$ is the maximum depth of the wetland. Volumetric evaporative flow rate ($Q_e$) out from the wetland depend on the evaporation rate calculated with the Penman equation ($e_p = e_{eq}$ for $r = 0$), and average area of the wetland during the period of groundwater inflow calculation $A(h)$

$$Q_e = e_p A(h)$$

(A11)

where $h$ is the average water level depth for the period of groundwater inflow calculation in cm and a polynomial relation was used to approximate the depth-area relation $A(h)$ based on detailed bathymetric surveys of the wetland.

The change in isotopic concentration is given by:

$$\frac{dC_i}{dt} = \frac{\Delta C_i}{\Delta t}$$

(A12)

$$\Delta C_i = C_i(t_f) - C_i(t_i)$$

(A13)

$$\Delta t = t_f - t_i$$

(A14)

Similar to the area of the wetland, the depth-volume relationship obtained from the bathymetric surveys was used to obtain a polynomial relation for $V(h)$. Specific groundwater inflow ($q_{GW}$) in units of $L/T$ was obtained by dividing the volumetric groundwater inflow by the transversal area of the wetland multiplied by the specific yield ($S_p$).

$$q_{GW} = \frac{Q_{GW}}{L/W/H_S}$$

(L15)

where $W$ is the total wetland set to 450 m and $S_p$ was set to 0.11 based on former studies at the site (Scholl et al., 2005). Groundwater outflow was calculated in a similar way.

References


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