A MODFLOW–based approach to simulating wetland–groundwater interactions in the Lower Limestone Coast Prescribed Wells Area

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Preface

South East Regional Water Balance Project Background

The South East Regional Water Balance project is a collaboration between Flinders University, CSIRO and the Department of Environment, Water and Natural Resources (DEWNR), funded by the Goyder Institute for Water Research. The project commenced in September 2012, with the objective of developing a regional water balance model for the Lower Limestone Coast Prescribed Wells Area (LLC PWA). The project was initiated following conclusions from the South East Water Science Review (2011) that, due to a number of gaps in understanding of processes that affect the regional water balance, there is uncertainty about the amount of water that can be extracted sustainably from the Lower Limestone Coast region as a whole. The review also concluded that, because of the close link between groundwater and surface water resources in the region, surface water resources and ecosystems are particularly vulnerable to groundwater exploitation.

The South East Regional Water Balance project follows on from the report of Harrington et al. (2011), which recommended that a consistent framework of models is required to support water management in the South East, with the first step being a regional groundwater flow model to:

- bring together all existing knowledge,
- address regional scale water balance questions
- provide boundary conditions for smaller scale models to address local scale questions, including those around "hotspot" areas and significant wetlands.

Harrington et al. (2011) also identified the critical knowledge gaps that limit the outcomes from a regional scale model. These included but were not limited to:

- Spatial and temporal variability in groundwater recharge and evapotranspiration.
- Interaquifer leakage and the influence of faults on groundwater flow.
- The nature of wetland-groundwater interactions
- Understanding of processes occurring at the coastal boundary
- Surface water-groundwater interactions around the man-made drainage network
- The absence of information on historical land use and groundwater extraction

The South East Regional Water Balance project has included numerous tasks that have sought to improve the conceptualisation of the regional water balance, address some of the critical knowledge gaps, incorporate this and existing information into a regional groundwater flow model and understand how this improved understanding can be used in the management of wetland water levels.

An overview of the project and its output can be found in Harrington et al. 2015. *South East Regional Water Balance Project – Phase 2. Project Summary Report.* Goyder Institute Report 15/39.

Associated publications

Technical Reports:

Harrington, N and Lamontagne, S (eds.), 2013*, Framework for a Regional Water Balance Model for the South Australian Limestone Coast Region.* Goyder Institute for Water Research Technical Report 13/14.

Morgan, LK, Harrington, N, Werner, AD, Hutson, JL, Woods, J and Knowling, M, 2015, *South East Regional Water Balance Project – Phase 2. Development of a Regional Groundwater Flow Model.* Goyder Institute for Water Research Technical Report 15/38.

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Executive Summary

Wetlands are defined as “those areas that are inundated or saturated by surface or groundwater at a frequency and duration sufficient to support, and that under normal circumstances do support, a prevalence of vegetation typically adapted for life in saturated soil conditions. Wetlands generally include swamps, marshes, bogs, and similar areas” (Tiner, 1996). From a hydrologists’ perspective, the functioning of a wetland may be characterised by the calculation of a water mass balance. The total water budget of a wetland may include components such as precipitation, open water evaporation, and transpiration by vegetation as well as surface water inputs and outputs*,* which may include environmental and/or managed flows. In areas where shallow water tables are present, such as the South East region, interactions with groundwater may be a significant additional component of a wetland water balance.

The aim of the present work was to develop an approach to simulating interactions between Lower Limestone Coast Prescribed Wells Area wetlands and underlying shallow groundwater; specifically, to translate water table variations near a wetland into changes in wetland water level. The industry standard groundwater flow simulation code MODFLOW (Hanson et al., 2014) was used as the basis for this approach. Methods of representing significant components of the water mass balances for both wetland and groundwater domains were assessed. For a wetland, these included precipitation on a wetland catchment area, evaporative losses from inundated areas and evapotranspiration losses from non-inundated areas, as well as additions and losses via both surface water and groundwater flows. For the groundwater domain, significant water mass balance components included lateral flows, including vertical leakage; evapotranspiration from shallow water tables; wetland–groundwater interactions; and changes in aquifer storage volumes. Each of these components was characterised as a time-varying flux. Of particular novelty was the combined approach used to represent recharge and evapotranspiration, which can be represented as a net flux from groundwater rather than by following the traditional approach of compartmentalising the two fluxes. Methods of extracting relevant information from model outputs were also addressed, including producing statistical summaries of wetland surface water persistence and calculating a simple salinisation risk metric in lieu of solute transport simulation.

The utility of the approach developed was demonstrated using a synthetic dataset, in lieu of outputs from the regional groundwater flow model for the region that is currently under development, which was unavailable for testing at the time of writing. Results of the synthetic demonstration, including the calculated salinisation risk metric, indicated the potential for managed surface water additions to negate the effects of long-term water table decline on the persistence of wetland surface water levels. Potential means of linking the wetland–groundwater interaction modelling approach to the regional groundwater flow model are also discussed. In future, the modelling approach described here could be used to identify the conditions which may lead to significant changes in wetland hydrological conditions, or to identify the timing of such changes for long-term variations in climate, land use and/or water allocation policy in the region.

1 Introduction

The *South East Water Balance Project* involves the development of tools to enable managers to assess the potential impacts of changes in climate, land use, and water allocation policy on groundwater resources and associated environmental assets in the Lower Limestone Coast Prescribed Wells Area (LLC PWA; referred to hereafter as “the South East region”). The main component of the project is the development of the first regional scale groundwater flow model for the region (Harrington and Lamontagne, 2013; Morgan et al., 2015). However, most LLC PWA wetlands cannot be represented in a regional scale model because they are smaller or of a similar size to individual model cells (~1 km^2). Thus, a tool is required to translate regional changes in the water table regime due to climate or other factors into possible changes in the water level regime of wetlands, a key ecological driver.

This report presents a generic, local scale wetland–groundwater interaction modelling approach for LLC PWA wetlands complementing the regional groundwater flow model. Some of the requirements considered in the development of the wetland–groundwater interaction model included that:

- Long term variations in the regional water table (whether obtained from the regional scale groundwater model or from other sources) could be incorporated;
- The approach is not too computationally demanding, thereby enabling the evaluation of a wide range of management scenarios over long time periods (i.e. decades to centuries);
- The tool is generic and applicable to deflation basin and other shallow (< 3 m) wetland types in the LLC PWA, rather than applicable to a specific wetland within the region;
- The output should be provided in the form of simple surface water level metrics that can be used by water managers and wetland ecologists to evaluate potential environmental impacts.

Interactions between wetlands and shallow groundwater are complex because they can be influenced by both regional and local factors. For LLC PWA wetlands, Harrington et al. (2015) hypothesised that six key regional and local scale factors influence surface water–groundwater interaction. These are:

1) The position of wetlands in the landscape (that is, whether located in a regional groundwater recharge, flow-through or discharge zone);
2) Landscape topography, e.g. the likelihood of local groundwater flow systems to develop in hummocky landscapes;
3) Subsurface controls on groundwater flow (e.g. geological basement intrusion), which can act to promote upward regional groundwater flow;
4) The presence or absence of clogging layers at the base of wetlands;
5) The use of surface water from drains or natural watercourses to supplement wetland water levels;
6) The morphometry of wetlands (including depth, surface area, and degree of incision in the landscape).

The approach to numerical modelling of wetland–groundwater interaction detailed in this report aims to address the effects of points 4, 5 and 6. However, the effects of landscape position and subsurface controls on the water table regime must be evaluated by other means, such as the regional scale groundwater flow model. A strategy to couple regional and local scale models is presented in the discussion of this report. However, at the time of writing the report, the regional scale groundwater flow model was still under development, so it was not possible to test this strategy. In the interim, in order to demonstrate the types of output that may be produced by a wetland–groundwater interaction model, a synthetic water table variation dataset was generated, which included a hypothetical five metre drop in the regional water table over a 100 year period.

2 Modelling approach

2.1 Conceptualisation

Wetlands may be defined as “those areas that are inundated or saturated by surface or groundwater at a frequency and duration sufficient to support, and that under normal circumstances do support, a prevalence of vegetation typically adapted for life in saturated soil conditions. Wetlands generally include swamps, marshes, bogs, and similar areas” (Tiner, 1996). Eleven classes of wetlands were defined by Cowardin et al. (1979), based upon factors such as the sediment composition of the wetland bed and shore as well as the dominant type of vegetation present. Wetlands located in the LLC PWA were classified into seven classes by Harding (2014), who described conceptual models for four of these classes: (1) inland interdunal wetlands and watercourses; (2) coastal dune lakes and permanent freshwater in drains; (3) karstic springs and coastal peat swamps; and (4) freshwater grass and sedge marshes. The hydrology of wetlands is typically highly variable (in terms of spatial location) as well as highly dynamic (in terms of temporal variation). From a hydrologists’ perspective, the functioning of a wetland may be characterised through calculation of a water mass balance. The total water budget of a wetland (Figure 1) may include components such as precipitation (*P*) and evapotranspiration (*ET*) fluxes as well as surface water inputs and outputs (*SWI, SWO*) which may include environmental and/or managed flows. In areas where shallow water tables are present, interactions with groundwater (*GWI, GWO*) may be a significant component of the wetland water balance.

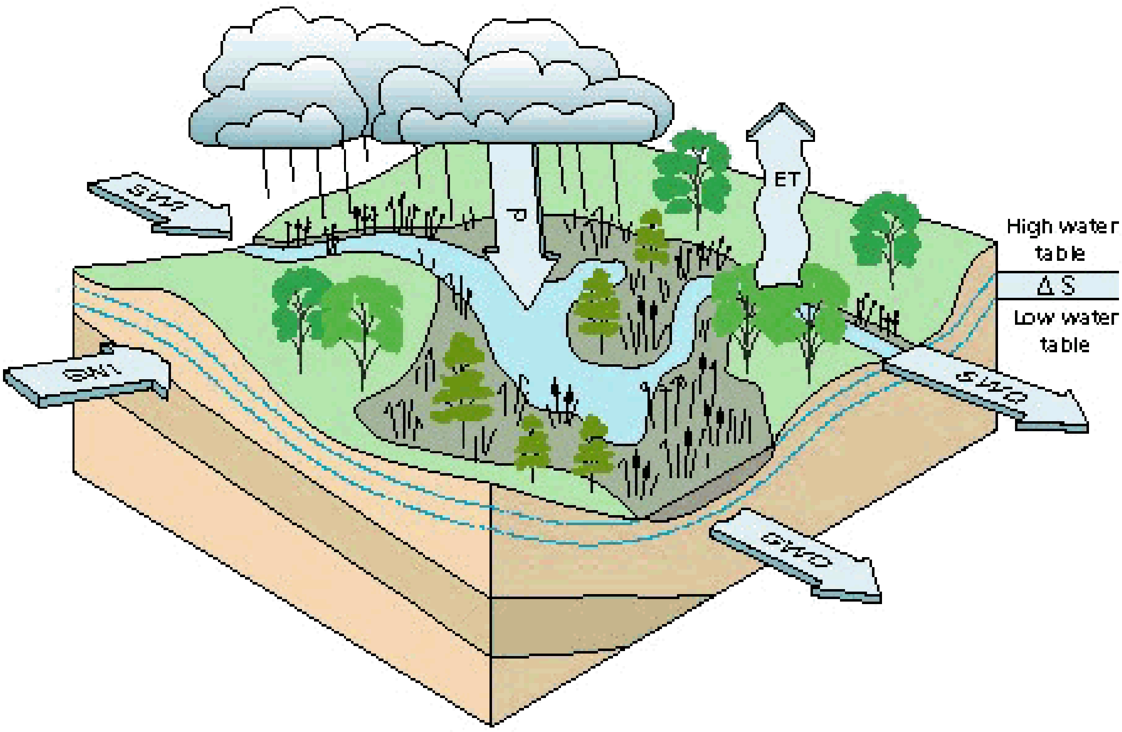

Figure 1. Three-dimensional conceptualisation of a wetland water budget where P = precipitation, SWI = surface-water inflow, SWO = surface-water outflow, GWI = groundwater inflow, GWO = ground-water outflow, ET = evapotranspiration and ΔS = change in storage (Carter, 1996).

The aim of the local scale wetland–groundwater modelling approach presented here is to translate water table variations near a wetland into changes in wetland water level. While three-dimensional simulations of wetland–groundwater interaction are not uncommon (Smerdon et al., 2007; 2012), many studies have successfully used two-dimensional cross sectional numerical models to provide valuable insights into the dynamics of surface water–groundwater interactions (Banks et al., 2011; Brunner et al., 2009a; Brunner et al., 2009b; Brunner et al., 2010; Crosbie et al., 2014; Shanafield et al., 2012). In addition, because this approach can represent physical processes efficiently, cross sectional models are able to simulate changes in wetland surface water regimes occurring over time periods of more than a century in length. For these reasons, a two-dimensional cross-sectional approach to simulating wetland–groundwater interactions was adopted in the present study. The model was designed to be applicable to deflation basins and other types of shallow wetlands (i.e., ≤3 metres depth) that are highly influenced by water tables variations as well as local scale features such as morphology and clogging layers. This is in contrast to wetlands that are associated with karst features (e.g., Blue Lake, Piccaninnie Ponds). These are largely a reflection of the regional water table since they are typically well connected to deep (i.e., >50 metres) groundwater flow paths and are not affected by capillary rise due to the absence of overlying low permeability sediments.

2.2 Model domain

The wetland–groundwater interaction model represents a cross-section of the landscape in a vertical plane oriented parallel to the direction of regional groundwater flow (Figure 2a). A wetland that is incised into the landscape is located in the centre of the model domain. The model domain features a total width w_{tot}, and the vertical extent of the model is defined as b_1 and b_2 at the left and right boundaries, respectively. The wetland features a width w_{WL} which may encompass a wetted (w_{wet}) and an exposed sediment area (w_{dry}); see Figure 3. Note that the latter two dimensions are not specified but are model outputs. The wetland may also have a clogging layer of known thickness (b_c). The distances from the wetland to the left and right model boundaries are specified as $w_{a(1)}$ and $w_{a(2)}$ respectively.

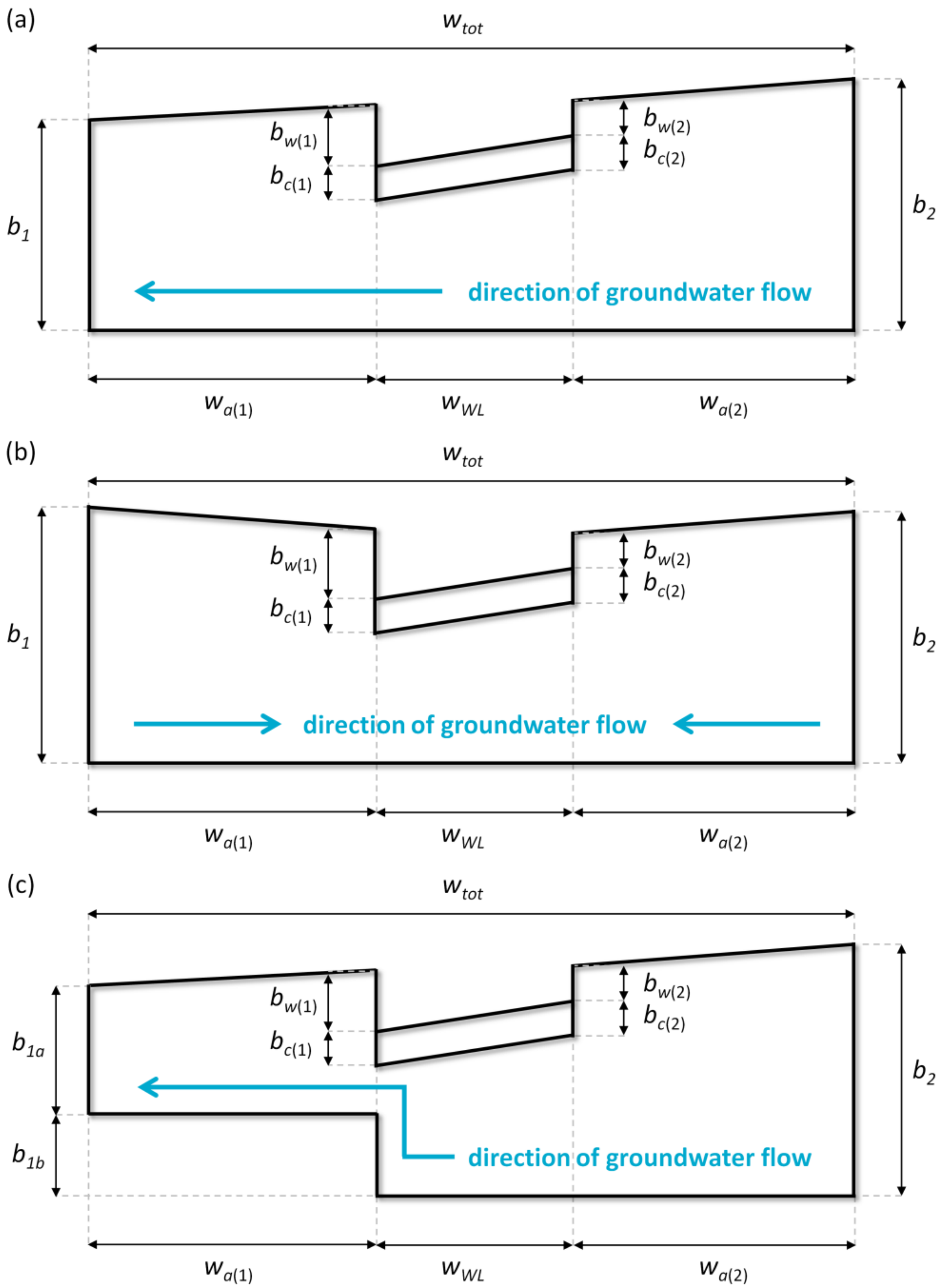

Figure 2. Hypothesised geometries of conceptual models of interaction between groundwater and a wetland located in (a) a regional flow-through zone, (b) a regional discharge zone and (c) a regional flow-through zone featuring bedrock intrusion. In order to improve clarity, topographic gradients are not to scale.

The dimensions of the model domain should be defined in terms of the regional and local scale hydrological environments to be represented. For example, when representing a wetland located in a regional flow-through environment, the value of b_2 will generally be greater than b_1 (Figure 2a). Alternatively, if a wetland is located in a regional discharge zone (e.g. between coastal dunes) then the value of b_2 could be equal to b_1 and greater than the elevation of the wetland (Figure 2b). In the case of regional flow-through wetlands, the vertical extent of the cross-section should be sufficient to enable the deepest (i.e. regional) groundwater flowpaths to bypass the wetland (that is, for the direction of groundwater flow to remain horizontal at depth). In some cases, such as when shallow bedrock intrusion is present, deep regional flow may be directed upward; the geometry of the model domain may be modified in order to represent this (Figure 2c).

Table 1. Symbols relating to model dimensions, as referred to by conceptual models of wetland–groundwater interaction (Figure 2a–c). All variables are in length units.

Symbol	Description
b_1	model vertical extent on left-hand boundary
b_2	model vertical extent on right-hand boundary
b_{1a}	aquifer vertical extent on left-hand boundary
b_{1b}	vertical extent of bedrock intrusion on left-hand boundary
$b_{c(1)}$	vertical extent of wetland clogging layer at left-most extent of wetland
$b_{c(2)}$	vertical extent of wetland clogging layer at right-most extent of wetland
$b_{w(1)}$	depth below land surface to which wetland is incised, at left-most extent
$b_{w(2)}$	depth below land surface to which wetland is incised, at right-most extent
$w_{a(1)}$	distance from left-most wetland extent to left-hand model boundary
$w_{a(2)}$	distance from left-most wetland extent to right-hand model boundary
w_{tot}	total width of model
w_{WL}	total width of wetland

The flow of water between a wetland and an underlying aquifer at any given time can be quantified by constructing water mass balances for each domain and by applying appropriate initial and boundary conditions. These two subjects are discussed in the following two sections, respectively.

2.3 Water mass balances

Water mass balances are based on the principle of conservation of mass; i.e., that the sum of inputs and outputs for a time-varying system will be equal to the change in mass stored in the system. Water mass balances are now described for both a wetland and an unconfined groundwater aquifer.

2.3.1 WETLAND WATER MASS BALANCE

At any given time, the various fluxes in and out of a wetland should be equal to the change in wetland storage volume over time, i.e.:

$$R_{net(\mathrm{wetland})}(t) + P(t) - E(t) + Q_{in}(t) - Q_{out}(t) + Q_{GW\rightarrow WL}(t) - Q_{WL\rightarrow GW}(t) = A_{WL}[h_{WL}(t)]\frac{\mathrm{d}h_{WL}}{\mathrm{d}t} \quad (1)$$

where each component is defined in Table 2. Note that the evaporation (E) and net recharge (R_{net}; i.e. net evapotranspiration) terms for the wetland are mutually exclusive. For inundated areas of the wetland it is assumed that water ponding results in plant death; consequently, the main surface flux is evaporation. Conversely, for dry areas of the wetland it is assumed that, where the watertable is sufficiently close to land surface, groundwater is lost to the atmosphere via evapotranspiration. When recharge is simultaneously taken into account, this is represented as a net recharge flux. In an area characterised by shallow depths to water table (such as the South East region), a key challenge is the accurate representation of water fluxes that occur at the land surface, i.e. open water evaporation and groundwater evapotranspiration. In the present work, these fluxes are represented using separate approaches, as discussed in Sections 2.4.3 and 2.4.4.

2.3.2 GROUNDWATER MASS BALANCE

At any given time, the various fluxes in and out of an unconfined aquifer should be equal to the change in the volume of groundwater storage over time, i.e. :

$$R_{net(\mathrm{land})}(t) \pm Q_1(t) \pm Q_2(t) \pm Q_3(t) - Q_{GW\to WL}(t) + Q_{WL\to GW}(t) = \frac{\mathrm{d}S}{\mathrm{d}t} \tag{2}$$

The components of the water mass balance for the combined wetland–groundwater system are listed in Table 2 and shown in Figure 3. It can be seen that governing equations (1) and (2) are linked by the two wetland–groundwater interaction flux terms $Q_{GW\to SW}$ and $Q_{SW\to GW}$.

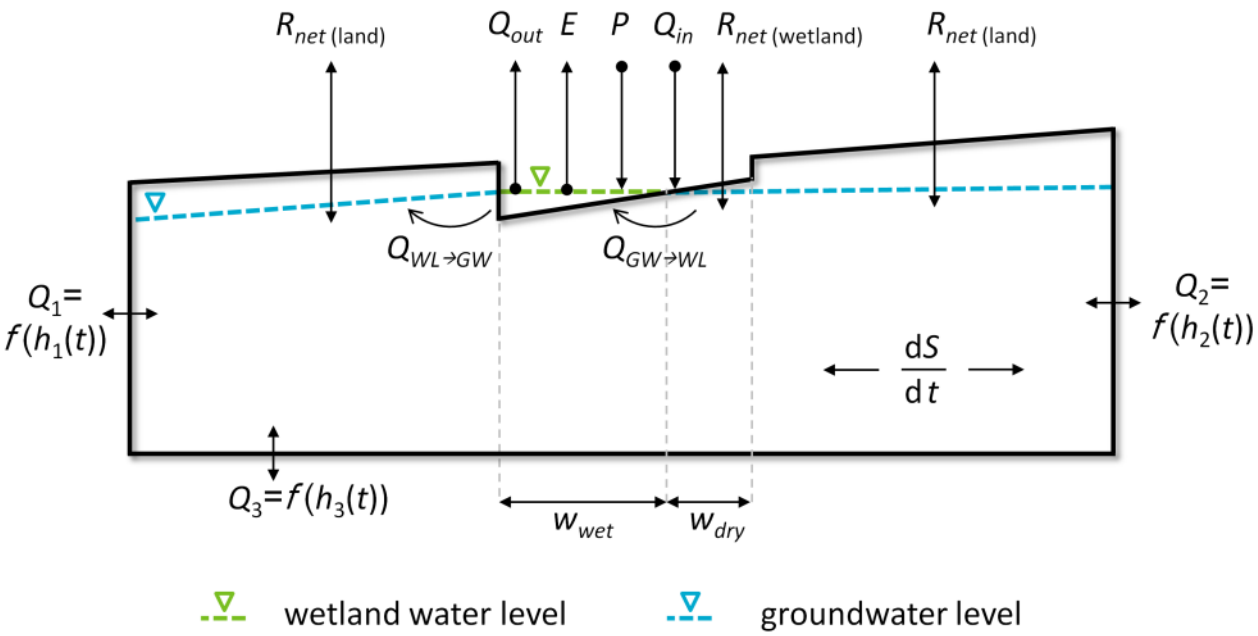

Figure 3. Water mass balance components (i.e. model boundary conditions) of the conceptual model of wetland–groundwater interactions in the South East region. Also shown are the groundwater level (blue dashes) and wetland water level (green dashes). In order to improve clarity, topographic and hydraulic gradients are neither consistent nor to scale. Symbols are described in Table 2.

Table 2. Water balance components (i.e. model boundary conditions) of the conceptual model of wetland–groundwater interaction.

Symbol	Description	Units
dS/dt	change in groundwater storage over time	$L^3.T^{-1}$
$E(t)$	wetland (open water) evaporation	$L^3.T^{-1}$
$P(t)$	precipitation on wetland	$L^3.T^{-1}$
$h_{GW}(t)$	groundwater hydraulic head	L
$h_1(t)$	hydraulic head on left-hand model boundary	$L^3.T^{-1}$
$h_2(t)$	hydraulic head on right-hand model boundary	$L^3.T^{-1}$
$h_3(t)$	hydraulic head on lower model boundary	$L^3.T^{-1}$
$Q_1(t)$	flux through left-hand model boundary	$L^3.T^{-1}$
$Q_2(t)$	flux through right-hand model boundary	$L^3.T^{-1}$
$Q_3(t)$	flux through lower model boundary	$L^3.T^{-1}$
$Q_{GW(in)}(t)$	regional groundwater flux into the conceptual model	$L^3.T^{-1}$
$Q_{GW(out)}(t)$	regional groundwater flux out of the conceptual model	$L^3.T^{-1}$
$Q_{GW\rightarrow WL}(t)$	flux from groundwater to wetland	$L^3.T^{-1}$
$Q_{in}(t)$	managed surface water addition to wetland	$L^3.T^{-1}$
$Q_{out}(t)$	managed surface water outflow from wetland	$L^3.T^{-1}$
$Q_{WL\rightarrow GW}(t)$	flux from wetland to groundwater	$L^3.T^{-1}$
$R_{net\,(land)}(t)$	net recharge to groundwater away from the wetland	$L^3.T^{-1}$
$R_{net\,(wetland)}(t)$	net recharge to groundwater below a dry wetland	$L^3.T^{-1}$
S_y	aquifer unconfined storage coefficient (i.e. specific yield)	-

2.3.3 GOVERNING EQUATION

Assuming two-dimensional groundwater flow in the vertical plane only, for a given representative elementary volume (Bear, 1972) of an unconfined aquifer this water mass balance may be rewritten as a partial differential equation, i.e. the 2-D governing equation for groundwater flow in an unconfined aquifer:

$$R_{net(\text{land})}(t) + K_x \frac{\partial^2 h}{\partial x^2} + K_z \frac{\partial^2 h}{\partial z^2} - Q_{GW\rightarrow SW}(t) + Q_{SW\rightarrow GW}(t) = S_y \frac{\partial h_{GW}}{\partial t} \qquad (3)$$

For given initial and boundary conditions, this governing equation can be solved using a standard numerical groundwater flow model to estimate the effects of fluctuating water tables on wetland surface water dynamics. Hydraulic parameters used in the solution of equation (3) are listed in Table 3.

Table 3. Hydraulic parameters used in the solution of the 2-D governing equation for flow in an unconfined aquifer.

Symbol	Description	Units
K_x	aquifer horizontal hydraulic conductivity	$L.T^{-1}$
K_z	aquifer vertical hydraulic conductivity	$L.T^{-1}$
K_C	wetland clogging layer hydraulic conductivity	$L.T^{-1}$
S_y	aquifer unconfined storage coefficient (i.e. specific yield)	-

Each water balance component listed in Table 2 will now be discussed in detail.

2.4 Boundary conditions

2.4.1 WETLAND–GROUNDWATER INTERACTION ($Q_{SW \rightarrow GW}$, $Q_{GW \rightarrow SW}$)

The movement of water between a wetland and an underlying aquifer includes infiltration (i.e. downward leakage) and exfiltration (i.e. upward discharge). Either one or both processes may occur at a given point in time. For example, wetlands located in regional recharge and discharge zones will primarily feature infiltration and exfiltration respectively. Conversely, a wetland located in a regional flow-through zone may feature significant contributions from both flux types. This regional scale conceptualisation of wetland–groundwater interaction is further complicated by local scale processes, including the presence or absence of a "clogging layer" at the base of a wetland. Such layers are typically composed of fine-grained sediments that are deposited as a result of flocculation and deposition. Clogging layers are typically characterised by a hydraulic conductivity that is significantly lower than that of an underlying aquifer. In addition, by retarding the flow of water between a wetland and an aquifer, the presence of clogging layers can lead to the occurrence of perched wetlands located above a water table.

In areas where a groundwater table does not intersect land surface, the vertical extent located between a wetland and groundwater is known as the vadose zone. Aquifer hydraulic conductivity, which governs the rate of water flow in the unsaturated zone, is dependent upon the degree of subsurface saturation. This relationship is described by the saturation-based form of Richards' equation for water flow in one dimension (i.e. vertical) under unsaturated conditions:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left[K_z(\theta)\left(\frac{\partial h}{\partial z} + 1\right)\right] \tag{4}$$

where z = elevation (L), K = aquifer hydraulic conductivity ($L.T^{-1}$), θ = water content (unitless), and h = aquifer hydraulic head (L). Richards' equation is inherently nonlinear, due to the dependence of aquifer hydraulic conductivity on pore water content. For this reason, solutions of Richards' equation are not trivial. Analytical solutions are limited while numerical solutions typically require fine spatial and temporal discretisation to ensure that solution convergence is achieved.

Alternatively, the vertical flow of water in the unsaturated zone between a wetland and a groundwater table can be approximated using a conductance-based approach. This method is derived from Darcy's Law for saturated water flow in porous media, i.e. :

$$q_z = K_z \frac{dh}{dz} \tag{5}$$

where q_z = linear water flux ($L.T^{-1}$), K_z = hydraulic conductivity ($L.T^{-1}$), dh = difference in hydraulic head (L) and dz = difference in position (L). The water flux in the unsaturated zone between a wetland and a groundwater table at time t can be approximated as:

$$q_{GW \leftrightarrow WL}(t) = q_z(t) = \frac{K_c}{b_c}\left[h_{WL}(t) - h_{GW}(t)\right] \tag{6}$$

which states that the flux is equal to the product of (1) the ratio of clogging layer hydraulic conductivity (K_c) to thickness (b_c) and (2) the hydraulic head gradient between the base of a wetland (h_{WL}) and an underlying aquifer (h_{GW}). Note that, in MODFLOW terminology, the term K_c / b_c is typically referred to as leakance (T^{-1}) (Harbaugh, 2005).

A volumetric flux is subsequently calculated by multiplying this linear flux by the lateral extent (in the horizontal plane) of the clogging layer.

Thus, the interaction of groundwater with wetlands may be represented using a hydraulic head–dependent specified flux (i.e. Cauchy) boundary condition, where the hydraulic gradient between a wetland and the underlying groundwater aquifer is calculated using the hydraulic heads computed for the previous stress period.

2.4.2 WETLAND SURFACE WATER INFLOW (Q_{in}) AND OUTFLOW (Q_{out})

Many wetlands in the South East region are supplemented by managed surface water inflows. These are managed through the South East drainage and floodway network using various control and diversion structures. The augmentation of storage volumes by managed surface water inflows is a significant component of the water mass balance for many South East wetlands. In addition, surface water inflows in the South East region typically occur during a brief period at the end of winter (Taylor et al., 2015). This mass balance component may be represented using a time-varying specified flux (i.e. Neumann) boundary condition. The episodic nature of surface water addition may be represented by specifying inflows at specified times only, e.g. :

$$Q_{in}(t) = \begin{cases} Q_{in\ (\text{specified})}, & t = t_{\text{specified}} \\ 0, & t \neq t_{\text{specified}} \end{cases} \tag{7}$$

This boundary condition serves to increase a wetland storage volume at a specified time.

Water balances for many South East wetlands include a surface water outflow component (i.e. Q_{out}). This can be represented simultaneously by specifying a time-varying specified flux boundary condition in a similar manner. Alternatively, where sufficient monitoring of both wetland inflows and outflows exist, a net flux may instead be specified.

2.4.3 WETLAND PRECIPITATION (P) AND EVAPORATION (E)

The contributions of precipitation and evaporation to a wetland water mass balance may be simulated as linear fluxes distributed over a nominated wetland area. These can be represented using time-varying specified flux (i.e. Neumann) boundary conditions. Linear fluxes of precipitation and evaporation may be based on local time series (e.g. rain gauge, Class-A pan) or interpolated time series (e.g. SILO datasets; https://www.longpaddock.qld.gov.au/silo/).

A volumetric precipitation influx is subsequently calculated by multiplying this linear flux by the lateral extent (in the horizontal plane) of the wetland.

In the case of evaporation data obtained from Class-A pan observations, it is worth noting that these will be greater than the true rate of evaporation from an open water body, such as a wetland. This is due to heating of the sides of a pan by the sun, which acts to promote additional evaporation. To account for this discrepancy between pan and open water evaporation, a coefficient ranging from 0.5 to 1.0 is typically applied to Class-A pan evaporation data (McMahon et al., 2013; Viessman et al., 1989), e.g. :

$$E_{wetland}(t) = k\ E_{pan}(t) \tag{8}$$

where $E_{wetland}$ and E_{pan} are volumetric fluxes (i.e. units of $L^3.T^{-1}$) and k is the evaporation pan coefficient.

A volumetric flux is subsequently calculated by multiplying this linear flux by the lateral extent (in the horizontal plane) of the inundated wetland area at time t.

2.4.4 NET RECHARGE ($R_{net\ (\text{wetland})}$, $R_{net\ (\text{land})}$)

Groundwater tables in the South East region are typically shallow (i.e. less than five metres below ground surface); for this reason, an accurate characterisation of evapotranspiration (ET) fluxes is critical. Recharge to and evapotranspiration from groundwater are often conceived as two independent mechanisms. Groundwater recharge is typically simulated as being independent of depth to water table; i.e. a specified flux (i.e. Neumann) boundary condition. Conversely, the evapotranspiration (ET) of water from groundwater is typically simulated as a hydraulic head–dependent specified flux (i.e. Cauchy) boundary condition. More specifically, ET is assumed to occur where the depth to groundwater is less than a nominated extinction depth. The rate of ET from groundwater is then defined as a function of depth, which, in its simplest form, decreases according to a linear or piecewise-linear trend from a nominated potential (i.e. maximum) ET value at ground surface to zero at the nominated extinction depth. This conceptualisation considers ET fluxes as being independent of recharge fluxes.

An alternative concept, net recharge, represents the net flux to groundwater when both recharge and ET fluxes are considered as being dependent upon depth to groundwater. In the South East region, net recharge values may vary from positive during winter months to negative during summer months. The present work uses this combined approach to represent recharge and ET for unsaturated areas of the wetland and for the adjoining land surface. Net recharge versus depth relationships have been generated for a range of soil, vegetation and climate combinations that are relevant to the South East region (Doble et al., 2015).

2.4.5 DOMAIN BOUNDARIES (Q_1, Q_2, Q_3)

Fluxes across the lateral model domain boundaries (i.e. Q_1, and Q_2) may be represented by time-varying hydraulic head (i.e. Dirichlet) boundary conditions (i.e. h_1 and h_2). Fluxes will be proportional to the hydraulic head gradients across these boundaries. The specification of transient hydraulic head values can enable the representation of temporal variations of varying periodicity, e.g. seasonal, decadal and/or longer term dynamics.

Fluxes across the lower model boundary (i.e. Q_3) may be represented by a specified flux (i.e. Neumann) boundary condition. Values should be specified in accordance with the regional and local hydrological context of the wetland of interest. In a regional scale context, a wetland may be located in a recharge, discharge or flow-through zone. For example, a wetland located in a regional recharge zone may be represented by imposing an influx along the lower boundary. Conversely, a wetland located in a regional discharge zone may be represented by imposing a discharge flux along the lower boundary. The parameterisation of vertical fluxes may be informed by outputs from the regional groundwater flow model (see Section 4 for further details) and/or from hydraulic, hydrochemical and/or isotopic testing.

2.4.6 GROUNDWATER STORAGE (dS/dt)

The storage and release of water in an unconfined aquifer, which is facilitated by water table fluctuation, is calculated as the product of aquifer specific yield and change in hydraulic head over time. Specific yield represents the volume of water released from an unconfined aquifer per unit surface area of the aquifer per unit decline in the water table (Freeze and Cherry, 1979). As described, this relationship can be used to represent changes in groundwater storage over time.

2.5 Numerical model

2.5.1 CHOICE OF MODELLING PLATFORM

The complexity of the conceptual model described, which features spatially variable hydraulic properties and geometry and temporal variations in boundary conditions, precluded the use of analytical solutions. Instead, a numerical modelling approach was used to simulate wetland–groundwater interaction.

The purpose of the local scale model is to translate water table variations near a wetland into changes in wetland water level. A number of numerical groundwater flow models have been used to simulate groundwater–wetland interactions, including MODFLOW (Harbaugh, 2005), HYDRUS (Simunek et al., 1999) and HydroGeoSphere (Therrien et al., 2006). The key difference between these modelling platforms is whether unsaturated zone processes (such as infiltration below perched wetlands) are represented explicitly by solving Richards' Equation for unsaturated flow (e.g. HYDRUS, HydroGeoSphere) or approximated through the use of a conductance term (e.g. MODFLOW). In a preliminary analysis, a number of groundwater flow models were tested for their suitability to simulate local scale wetland–groundwater interactions. Models based on solutions to Richards' Equation computed using HYDRUS were found to be less tractable due to difficulties in achieving numerical convergence under non-ideal conditions (such as non-loam soils; C. Turnadge, unpublished data). Conversely, models featuring conductance-based approximations of wetland–groundwater interaction were found to be more tractable at a broad range of spatial and temporal scales. The computational requirements of such models are also relatively low. For these reasons, MODFLOW was selected to simulate the conceptual model described above. In addition, the MODFLOW platform is widely used by the future anticipated end-users of the model (i.e. DEWNR staff) and is also the platform used for the regional LLC PWA groundwater flow model.

Specifically, the MODFLOW-OWHM (Hanson et al., 2014) version of MODFLOW was chosen in preference to the industry standard version (i.e. MODFLOW-2005 ; Harbaugh 2005). This was because it features additional capabilities for the simulation of environments featuring shallow water tables. In particular, MODFLOW-OWHM includes the Riparian-Evapotranspiration (RIP-ET) package (Baird and Maddock, 2005; Maddock et al., 2012), which enables the inclusion of non-monotonic ET versus depth functions and the representation of zero ET for fully inundated vegetation. MODFLOW-OWHM is fully backwards compatible with MODFLOW-2005 and the majority of MODFLOW-OWHM packages are consistent with the previous release. In addition, many standard MODFLOW packages were re-coded for the MODFLOW-OWHM release to achieve improved numerical stability and convergence (W. Schmid, pers. comm., 26 November 2014).

Unlike the regional groundwater flow model, which is a specific hydrogeological model of the South East region, the wetland–groundwater interaction model described here is generic in nature. It does not aim to represent a specific wetland in the LLC PWA area. Although a demonstration model is provided later in this report, what is provided in the current section (2.5) is a series of instructions for how to set up such a model using MODFLOW–OWHM. Based on field studies of three wetlands in the region (Smith et al., 2015; Taylor et al., 2015) and a literature review, realistic parameter ranges are provided in order to guide future application of the model. Sources of data required for the model (e.g. precipitation, evaporation) are also provided.

2.5.2 LIMITATIONS OF MODELLING APPROACH

A key limitation of the use of saturated groundwater flow models to simulate wetland–groundwater interactions is the inability to account for unsaturated zone processes. For example, when the water level in a wetland reduces below the base of the wetland, the pressure head at wetland base will become negative and the hydraulic head will reduce accordingly. When this scenario is simulated using a saturated groundwater flow model such as MODFLOW, the minimum pressure head at the wetland base will be set to the elevation of the wetland base. Associated hydraulic head values will therefore be underestimated, as will the rate of flow through the wetland base. Therefore, the modelling approach presented here is a compromise between (a) generality and ease of use and (b) correct representation of the physics of unsaturated subsurface water flow. Since the presence and characteristics of clogging layers for wetlands in the South East region are not well known, the approach presented here is considered sufficient at present. Brunner et al. (2010) provide further advice for determining when the approach described here is sufficient to characterise unsaturated zone processes below wetlands.

2.5.3 IMPLEMENTATION OF CONCEPTUAL MODEL

With regards to MODFLOW packages, this conceptual model can be implemented as follows.

The Discretisation (DIS) package can be used to specify the spatial and temporal discretisation of the model. Input data for this package (such as geometry) could potentially be sourced from the regional groundwater flow model; this is discussed at length in Section 4. The Layer Property Flow (LPF) package can be used to specify the hydraulic properties described in Section 2.3.3.

The Constant Hydraulic Head (CHD) package can be used to specify time-varying boundary conditions for groundwater hydraulic heads on lateral model boundaries. Input data for this package could potentially be sourced from the regional groundwater flow model; this is discussed at length in Section 4.

The Lake (LAK) package can be used to specify boundary conditions for wetland precipitation, evaporation and surface water inflows and outflows, as well as clogging layer properties and interactions with groundwater. Input data for climatic boundary conditions may be sourced, for example, from the SILO meteorological database (https://www.longpaddock.qld.gov.au/silo/). Input data for boundary conditions representing surface water addition may be sourced from the WaterConnect database (https://www.waterconnect.sa.gov.au/Systems/SWD/).

The Riparian–Evapotranspiration (RIP–ET) package (Baird and Maddock, 2005; Maddock et al., 2012) can be used to represent net recharge to groundwater (i.e. $R_{net(land)}$). This package is particularly suited to representing nonlinear flux versus water table elevation relationships and can be adapted in order to represent net recharge (rather than ET fluxes alone). Separate flux versus depth relationships are required to represent the net recharge occurring from (1) cells underlying wetland (i.e. LAK package) cells and (2) cells located in adjoining land surface areas. The key difference between these two relationships is the flux calculated when the groundwater table elevation is above the elevation of the cell top (i.e. when the cell is fully saturated). For areas adjoining the wetland, this circumstance represents the ponding of water at the land surface and the net flux will be equal to potential evapotranspiration (i.e. evaporation), as observed in net recharge relationships derived from Richards' equation models (Doble et al., 2015). In comparison, the net recharge occurring from inundated cells underlying LAK packages cells will be zero, as net fluxes at the land surface boundary will be calculated by the LAK package.

The distinction between these two approaches is illustrated further as follows (Figure 4). Consider two columns of a model, the uppermost cells of which represent a wetland. The cell located in layer 1, column 1 is inundated by surface water; therefore precipitation, evaporation and surface water addition fluxes will be applied using the LAK package. The net flux between this cell and the cell in the layer below (i.e. the net flux to groundwater) will also be calculated. Net recharge to the underlying cell will not be calculated by the RIP–ET package because the LAK cell is inundated.

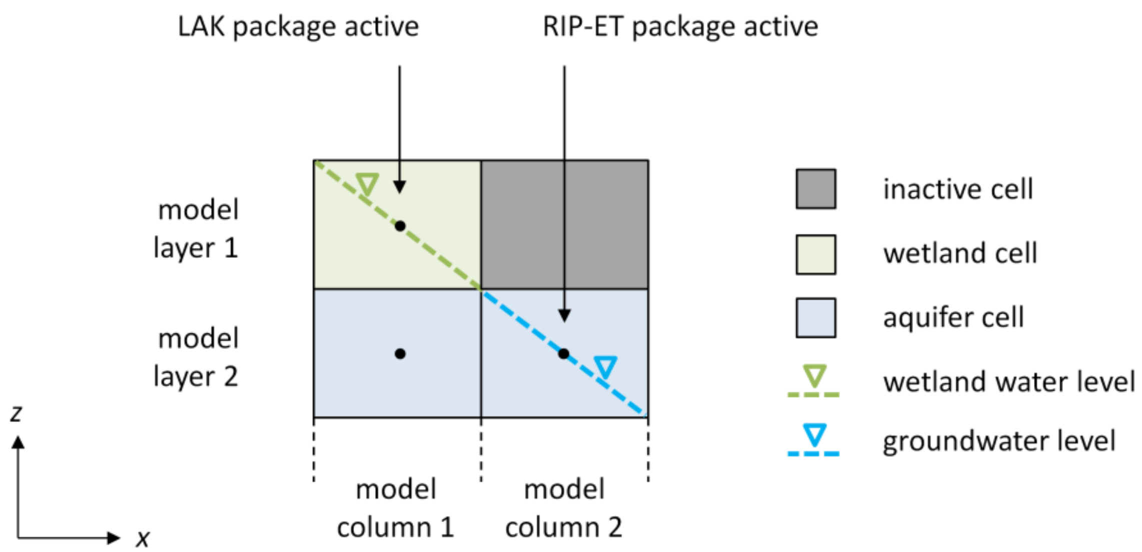

Figure 4. Example of the combined use of Lake (LAK) and Riparian–Evapotranspiration (RIP–ET) packages to represent precipitation, evaporation, surface water input, recharge, and evapotranspiration fluxes for a wetland atop an unconfined aquifer using MODFLOW–OWHM.

In comparison, the cell located in layer 1, column 2 is not inundated by surface water; therefore no fluxes will be applied using the LAK package and the net flux to groundwater will not be calculated. Net recharge to the underlying cell will be calculated using the RIP–ET. In practice, due to the use of transient boundary conditions on the lateral boundaries of the model, the degree of inundation of wetland cells will vary with time.

The transition from inundated conditions (and LAK package fluxes) to unsaturated conditions (and RIP–ET package fluxes) will therefore be time–variant. With regards to the drying and rewetting of cells, which is a common problem in MODFLOW models (Doherty, 2001), this problem can be avoided by using a sufficiently thick model layer to represent the uppermost vertical extent of an unconfined aquifer. For example, if a groundwater flow system of interest features water table variations of less than 10 metres in magnitude then a 10 metre thick model layer could be employed. This approach permits significant variation of water table elevations without leading to the drying of cells.

The RIP–ET package, however, does not allow for a non-zero value to be specified at the lower extent of the specified flux versus depth relationship. In order to circumvent this limitation, for a given net recharge versus depth function, the maximum depth (and associated net recharge value) of the relationship can be extended to greater depth; for example, to the base of the relevant model layer.

2.5.4 POST-PROCESSING OF MODEL OUTPUTS

Outputs produced by a transient MODFLOW model typically include time series of hydraulic head and flow velocity on a cell-by-cell basis. These raw outputs may be post-processed in order to identify results that are salient to the simulated problem of interest. For the simulation of wetland–groundwater interactions, three results are of particular interest:

1. **Time series of hydraulic head for the cell representing the lowest wetland elevation.** This can be used to represent changes in surface water level for the entire wetland, regardless of bathymetry.
2. **Box and whisker plots of wetland surface water level regime metrics** (e.g. the minimum and maximum annual fraction of the wetland inundated). Box and whisker plots can be used to summarise the mean, median, minima, and maxima of time series data. The may be used to summarise a wetland surface water level regime for a given time period (e.g. a 100 year period).
3. **A wetland salinity risk index.** For a given wetland, a simple index may be calculated by which to estimate the risk of salinisation. Here we define the salinity index (*SI*) as the ratio of wetland evaporation (*E*) as a fraction of the total outflux (for a wetland that does not feature managed surface water outflows), i.e. :

$$SI = \frac{E}{E + Q_{SW \to GW}} \tag{9}$$

Outputs from the MODFLOW model were defined using the Output Control (OC) and Gauge (GAGE) packages. See Appendices A.8 and A.9 for example OC and GAGE package input files respectively. Post-processing of MODFLOW outputs may be undertaken using a scripting language. For the demonstration example presented here, Python language scripts were written and are provided in Appendices C.1 and C.2.

3 Model demonstration using a synthetic dataset

3.1 Conceptual model

As discussed previously (Section 1), as the regional scale LLC PWA groundwater flow model was incomplete at the time of writing, a hypothetical scenario was instead developed to demonstrate the utility of the wetland–groundwater interaction simulation model. A wetland located in a regional flow-through zone was simulated using a 5000 m wide domain that varied in thickness from 60 m to 64 m (Figure 5). The wetland feature represented was 1000 m in width and featured a depth of 2 m below land surface; this is considered to be a common morphometry for LLC PWA deflation basin wetlands (Taylor et al., 2015). The effects of the presence and absence of a one metre-thick clogging layer were tested. While located in a regional flow-through zone, three wetland types were considered: (1) a local scale recharge wetland; (2) a local scale flow-through wetland; and (3) a local scale discharge wetland. In addition to clogging layer presence and local scale connection status, the effects of managed wetland surface water inputs of various magnitudes were also examined.

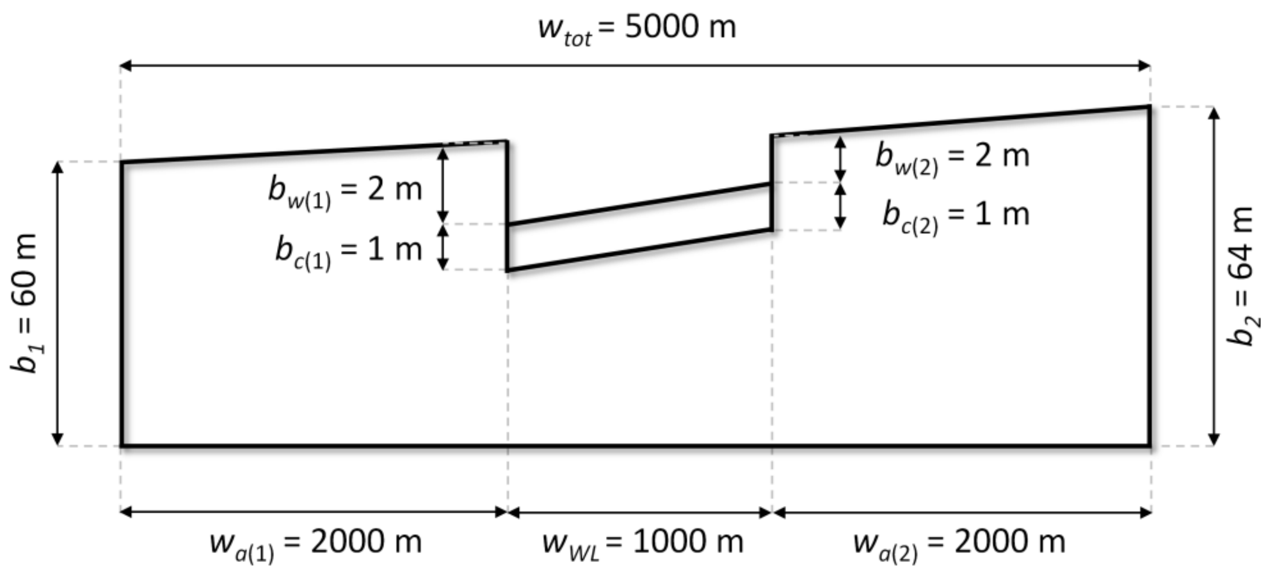

Figure 5. Conceptual model geometry for synthetic dataset-based demonstration example.

3.2 Model domain and discretisation

In order to solve the 2-D unconfined groundwater flow equation, the conceptual model domain was discretised spatially using 100 metre-wide cells, including the area representing the wetland feature (Figure 6). Such coarse horizontal discretisation can be justified if it can be assumed that fluxes between groundwater and the wetland are primarily vertical. The vertical extent of the numerical model was discretised using six model layers. Model layers four to six featured a uniform thickness of 10 metres. The thickness of layer three was varied from 10 m at the left-hand boundary to 14 m at the right-hand boundary. The thickness of layers one and two was a uniform 10 m except below the wetland, where it was 2 m and 18 m respectively.

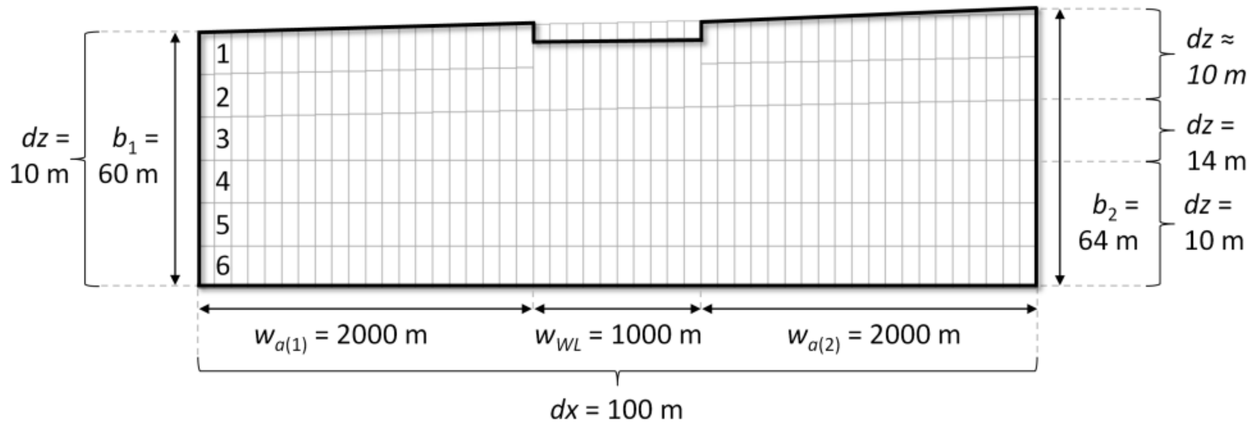

Figure 6. Spatial discretisation of conceptual model for synthetic dataset-based demonstration example, including model layer numbers (note: vertical scale is exaggerated for clarity).

In order to simulate water table decline over periods of hundreds of years, a 400 year-long temporal extent was specified. This was composed of four 100 year-long periods: (a) a model warm-up period, to allow for pseudo-equilibration of initial and boundary conditions; (b) a pre-water table decline period of semi-equilibrium; (c) a period featuring a water table decline of five metres; and (d) a post-water table decline period of semi-equilibrium.

The 400 years of simulated time was discretised using 30 day-long stress periods, each of which was subdivided into 30 time steps; a stress period length multiplier was not used as specified fluxes were constant within stress periods. The use of 30 day-long periods to approximate months was chosen primarily to achieve consistency with the temporal resolution of the regional scale groundwater flow model. In addition, the use of sub-annual discretisation allowed for the inclusion of seasonal dynamics in transient domain boundary conditions (see Section 3.4.5).

3.3 Model parameterisation

The groundwater aquifer represented in the demonstration model features a homogeneous isotropic hydraulic conductivity of 5 $m.d^{-1}$, which is consistent with that of a karstic limestone aquifer (Freeze and Cherry, 1979). The equivalent transmissivity of a 60 m thick aquifer (i.e. 300 $m^2.d^{-1}$) is consistent with the range reported by Harrington and Lamontagne (2013) for the Tertiary Limestone Aquifer (i.e. 100-400 $m^2.d^{-1}$), the primary regional aquifer in the South East region. Aquifer specific yield and specific storage values were specified as 0.1 and 5 x 10^{-4} m^{-1} respectively. (Although an unconfined aquifer was simulated, due to vertical discretisation of the aquifer into a number of layers, many layers remained fully saturated throughout the duration of the model and therefore required use of a confined storage coefficient.) Model parameter values are summarised in Table 4.

Table 4. Parameters values specified for the synthetic dataset-based demonstration example.

Symbol	Description	Value(s)	Units
b_1	aquifer thickness on left-hand model boundary	60	m
b_2	aquifer thickness on right-hand model boundary	64	m
b_C	thickness of wetland clogging layer (uniform)	1	m
dh_{WL}	change in wetland stage due to surface water addition	0.0, 0.5, 1.0, 1.5	m
E	wetland evaporation	$f(t)$	$m.d^{-1}$
$ET_{net\ (land)}$	evapotranspiration from groundwater away from wetland	$f(t)$	$m.d^{-1}$
$ET_{net\ (wetland)}$	wetland evapotranspiration	$f(t)$	$m.d^{-1}$
$h_1\ (t=0)$	initial groundwater hydraulic head on left-hand boundary	54	m
$h_2\ (t=0)$	initial groundwater hydraulic head on right-hand boundary	58	m
K_x	aquifer horizontal hydraulic conductivity	5	$m.d^{-1}$
K_z	aquifer vertical hydraulic conductivity	5	$m.d^{-1}$
K_C	wetland clogging layer hydraulic conductivity	0.227	$m.d^{-1}$
P	wetland precipitation	$f(t)$	$m.d^{-1}$
S_s*	aquifer confined storage coefficient (i.e. specific storage)	5×10^{-3}	m^{-1}
S_y	aquifer unconfined storage coefficient (i.e. specific yield)	0.1	-
w_{tot}	total model width	5000	m

*Note: Sub-discretisation of an unconfined aquifer in the vertical plane results in fully saturated model cells; hence, numerical solutions of groundwater flow in such cells require the use of a saturated storage coefficient.

The Discretisation (DIS) package was used to specify the spatial and temporal discretisation of the model. See Appendix A.1 for an example DIS package file. The Layer Property Flow (LPF) package was used to specify values for hydraulic properties; see Appendix A.2 for an example LPF package file.

3.4 Initial and boundary conditions

Initial and boundary conditions specified for the demonstration model are now described.

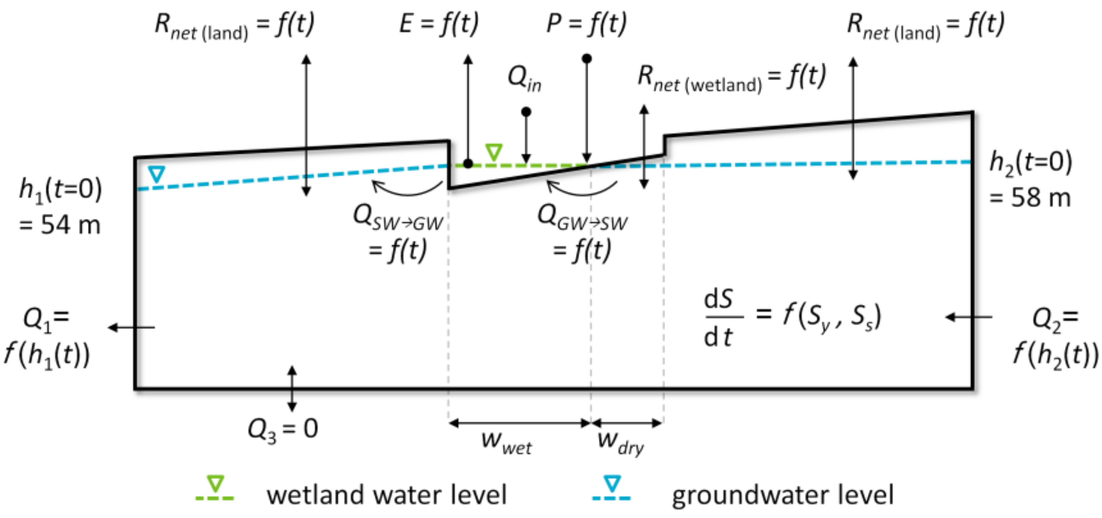

Figure 7. Water mass balance components (i.e. boundary conditions) for synthetic dataset-based demonstration example.

3.4.1 WETLAND–GROUNDWATER INTERACTION ($Q_{SW \rightarrow GW}$, $Q_{GW \rightarrow SW}$)

To assist the characterisation of wetland–groundwater connectivity in the South East region, various field studies including soil and environmental tracer sampling (Smith et al., 2015; Taylor et al., 2015) were undertaken in addition to a desktop study of available hydrological data. A key outcome of the soil sampling undertaken was the characterisation of the clogging layer present at Bool Lagoon, including its spatial extent and hydraulic properties. In terms of informing the conductance–based approach to estimating wetland–groundwater interaction described in Section 2.4.1, these data provided a physical basis for specification of clogging layer hydraulic conductivity and thickness. From a set of 15 samples distributed across Bool Lagoon, the mean saturated hydraulic conductivity of the wetland clogging layer was calculated as 0.227 m.d^{-1} (Taylor et al., 2015), which is consistent with a silt porous medium (Freeze and Cherry, 1979). A mean clogging layer thicknesses of eight metres was calculated from lithological log analysis. Some logs indicated a clogging layer thickness in excess of ten metres. For simplicity, however, in the present modelling study the thickness of the simulated clogging layer was assumed to be one metre. Since clogging layer leakance is defined as the ratio of hydraulic conductivity to thickness, a value of 0.227 d^{-1} was used.

3.4.2 WETLAND SURFACE WATER AUGMENTATION (Q_{in})

The surface water level regime of many wetlands of the South East region is manipulated through the use of regulators and flows via constructed drains. Four surface water augmentation scenarios were considered in the present study. Specified inflows were equal to volumes sufficient to fill a dry wetland to a stage level (i.e. $h_{WL\,(target)}$) of 0.0, 0.5, 1.0, or 1.5 m. Increases in wetland stage due to managed surface water inflows were assumed to occur in September of each year, which is consistent with the management of LLC PWA wetlands such as Bool Lagoon.

$$Q_{in}(month) = \begin{cases} h_{WL\,(target)} * A\,, & \text{month} = \text{September} \\ 0.0\ \text{m}^3, & \text{month} \neq \text{September} \end{cases} \tag{10}$$

where A = cell area. For the demonstration model presented here, which does not feature a wetland outflow component, the Lake package parameter WTHDRW was used to specify volumetric surface water additions. For a model featuring both surface water additions and outflows, the Lake package parameters RNF and WTHDRW could be used to simultaneously represent additions and outflows respectively.

3.4.3 WETLAND PRECIPITATION (*P*) AND EVAPORATION (*E*)

In order to simulate wetland precipitation and evapotranspiration fluxes appropriate to the South East region, climate data for the period 1911–2010 and interpolated to a point located in Bool Lagoon (approximately 37.1°S, 140.7°E) were obtained from the Bureau of Meteorology SILO database (https://www.longpaddock.qld.gov.au/silo/). These daily data were aggregated to monthly resolution and mean values were calculated for each month of the calendar year (Figure 8). A sinusoidal function was fitted to the resulting values for monthly wetland precipitation:

$$P(t) = -30\sin\left[-1.7\pi\left(\frac{t-1}{11}\right) + 1.5\right] + 50 \tag{11}$$

where P is wetland precipitation (mm.month^{-1}) and t is total model time elapsed in months (an integer). Similarly, a sinusoidal function was fitted to aggregated Class-A pan evaporation data for Bool Lagoon:

$$E(t) = 0.97\left\{100\sin\left[-1.8\pi\left(\frac{t-1}{11}\right) + 1.3\right] + 140\right\} \tag{12}$$

where E is wetland evaporation (mm.month^{-1}) and t is total model time elapsed in months (an integer).

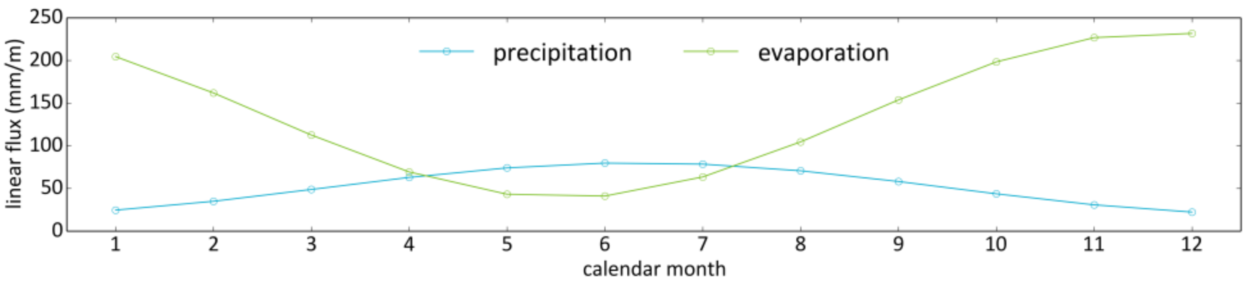

Figure 8. Mean intra-annual dynamics for wetland precipitation and evaporation fluxes, as obtained from interpolated climate data for the period 1911-2010.

The value of 0.97 used in the latter equation is a pan correction factor, which is commonly used to account for hydrodynamic effects when converting evaporation measured using a small Class-A pan to evaporation from a large surface water body (McMahon et al., 2013; Viessman et al., 1989). The value used is consistent with the mean annual pan correction coefficient calculated from Mount Gambier aerodrome (McMahon et al., 2013). Other coefficients used in Equations 11 and 12 were estimated by calibrating these functions by hand to the aggregated climate data. The precipitation and pan-corrected evaporation functions presented are consistent with annual rates of approximately 630 mm.y^{-1}and 1600 mm.y^{-1} respectively.

See Appendix A.6 for an example LAK package file, which includes the representation of wetland–groundwater, precipitation and evaporation fluxes as well as surface water additions. See Appendix B.4 for the Python language script used to automate the preparation of LAK package input files.

3.4.4 NET RECHARGE ($R_{net\,(\mathrm{wetland})}$, $R_{net\,(\mathrm{land})}$)

MODFLOW groundwater flow models typically simulate evapotranspiration from groundwater using either the Evapotranspiration (EVT) package or the Evapotranspiration Segments (ETS) package. Both packages simulate ET as a function of depth below ground surface when the water table is above a nominated extinction depth. A monotonic decline in ET from potential ET (at land surface) to zero (at a given extinction depth) can be simulated using a linear function (EVT package) or piecewise linear function (ETS package).

In the present work evapotranspiration from groundwater was instead represented using the Riparian Evapotranspiration (RIP-ET) package (Maddock et al., 2012). Use of this package provides three key advantages over "traditional" methods of ET representation. First, the RIP-ET package can be used to represent negative values of evapotranspiration (i.e. net recharge). Second, the RIP-ET package is suitable for the simulation of nonlinear, non-monotonic ET versus depth relationships. These are represented using a piecewise linear approach. Net recharge versus depth relationships are often non-monotonic; for example, maximum ET may occur below ground surface, indicating plant rooting depth.

Third, the procedure by which the ET versus depth relationship is specified allows for a zero flux to be specified when the water table is at (or immediately above) a given elevation. This applies particularly to model cells located directly below a Lake package cell. When the watertable is in a Lake package cell, a zero ET flux will be applied to the cell below. If the watertable declines below the base of the Lake package cell then ET will be removed from the cell below, until the watertable declines below the nominated extinction depth.

As part of the South East Water Balance Project, Doble et al. (2015) generated a large number of net recharge versus depth functions for a range of soil and land use types, as well as for a range of latitudinal locations relevant to the South East region. The inclusion of multiple locations along a latitudinal gradient was used to account for climatic variability. In order to demonstrate the ability to link the present work with that of Doble et al. (2015), a single net recharge function was adopted (Figure 9). This function was generated for a silty soil (with an extinction depth of ~5 m) featuring an annual vegetation type and located at a similar latitude to that of Bool Lagoon. For the purposes of demonstration modelling, this function represents the mean annual variability in net recharge. In practice, this relationship would vary with time, particularly reflecting seasonality. Two functions are shown in Figure 9; these differ by the flux applied when the water table is above the top of the model cell (i.e. when the depth to water table is less than zero). For model cells located below a wetland (i.e. Lake package) cell, this flux is equal to zero (Figure 9a). Conversely, for model cells located adjacent to wetland areas, this flux is equal to potential evapotranspiration (Figure 9b). See Appendix A.5 for an example RIP-ET package file.

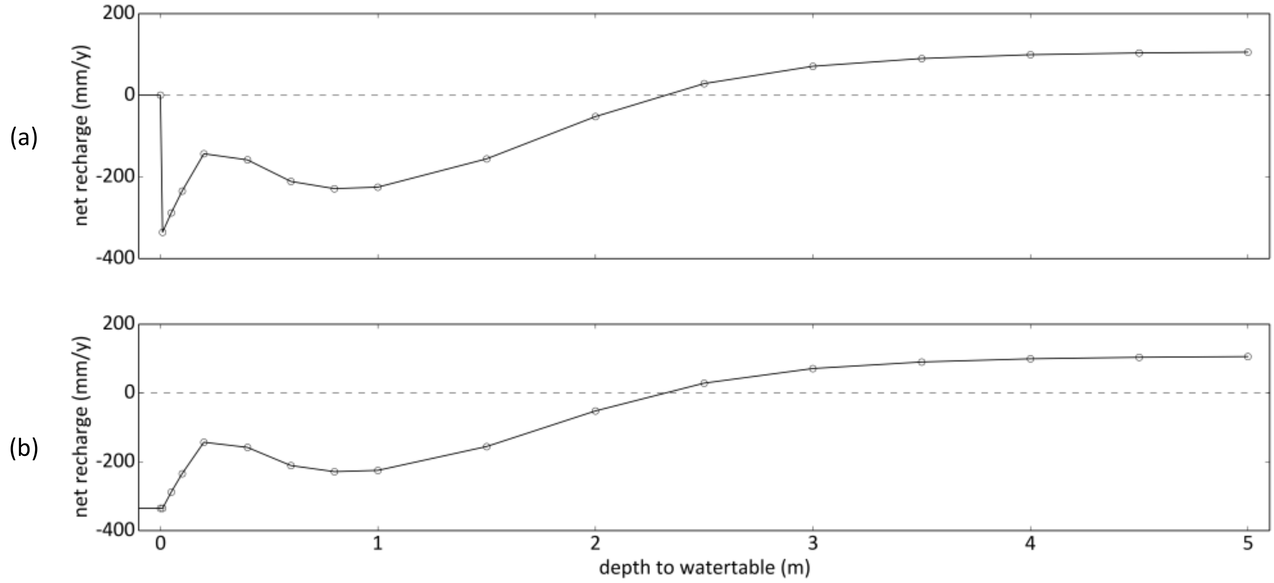

Figure 9. Net recharge flux versus depth relationships for (a) non-inundated wetland areas and (b) adjacent to wetland areas.

3.4.5 DOMAIN BOUNDARIES (Q_1, Q_2, Q_3)

Initial conditions along lateral and lower model boundaries were specified as appropriate for a regional flow-through wetland type. A hydraulic head of 58 m was specified along the outflow lateral boundary, corresponding to a water table elevation located at the top of each wetland (i.e. LAK package) cell. A hydraulic head of 62 m was specified along the inflow lateral boundary; the corresponding hydraulic gradient is therefore 4 m / 5000 m, = 8×10^{-4}, which is consistent with the typical magnitude of a groundwater hydraulic gradient. A zero flux was specified across the lower model boundary, which is consistent with shallow groundwater flow in the vicinity of a flow-through-type wetland. Long term equilibrium of groundwater conditions was achieved by running the model for the duration of the extra initial 100 year-long period.

In order to simulate water table decline over periods of hundreds of years, a synthetic time series of hydraulic head was developed using the linear sum of two component series. These were: (1) century-scale groundwater dynamics, including water table decline; (2) decadal-scale climate dynamics of drought and flood; and (3) intra-annual climate seasonality, i.e.:

$$h_{GW}(t) = h_{GW_1}(t) + h_{GW_2}(t) + h_{GW_3}(t) \tag{13}$$

Century-scale groundwater dynamics were specified using three periods: (a) a 200 year-long pre-decline period of semi-equilibrium; (b) a 100 year-long period featuring a water table decline of five metres; and (c) a 100 year-long post-decline period of semi-equilibrium, i.e.:

$$h_{GW_1}(t) = \begin{cases} h_{t=0}, & 0 < t \leq 200 \\ h_{t=200} - (5/100)\, t, & 201 < t \leq 300 \\ h_{t=300}, & 301 < t \leq 400 \end{cases} \tag{14}$$

where t = time elapsed (years) (Figure 10). A watertable decline of five metres was chosen arbitrarily but is considered to be a realistic groundwater response to climatic variation in the South East region.

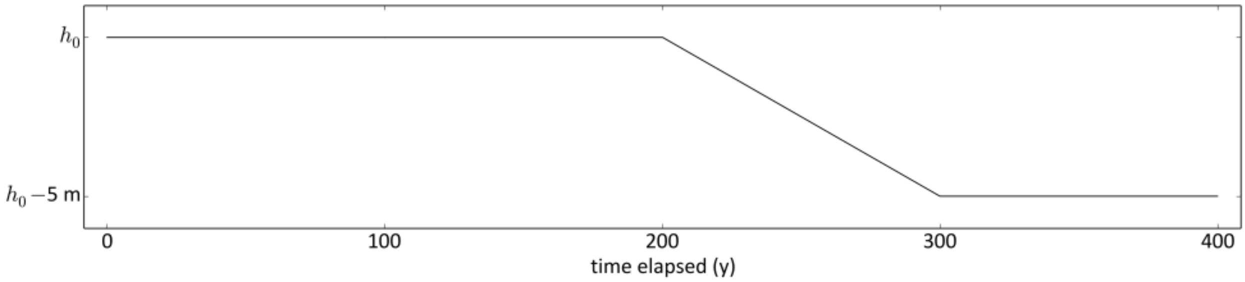

Figure 10. Long term (i.e. century scale) hypothetical watertable dynamics, including (a) pre-decline equilibrium period, (b) water table decline, and (c) post-decline equilibrium period.

Decadal-scale climate dynamics were represented using the sinusoidal function:

$$h_{\mathrm{GW}_2}(t) = 0.75 \sin\left(\frac{t}{12} * \frac{2\pi}{50}\right) \tag{15}$$

where t = time elapsed (months). The parameters of this function were specified in order to achieve a peak-to-peak period of 50 years, representing two wet periods and two dry periods over a 100 year-long period (Figure 11).

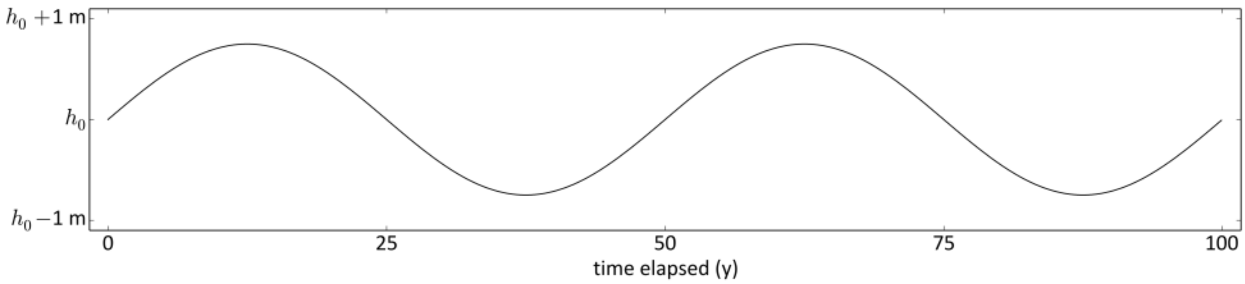

Figure 11. Medium term (i.e. decadal scale) hypothetical variations in regional watertable elevation induced by wet and dry cycles.

When included in the linear sum of component time series, this function features a peak-to-peak amplitude of approximately two metres, which is consistent with decadal-scale climate dynamics.

Intra-annual seasonal variation was represented using a second sinusoidal function, with a wave peak occurring in October and a nadir occurring in April:

$$h_{\mathrm{GW}_3}(t) = 0.25 \sin\left[2\pi\left(\frac{t}{12} + \frac{6}{12}\right)\right] \tag{16}$$

where t = time elapsed (months). The parameters of this function were specified in order to achieve a peak-to-peak period of one year (Figure 12).

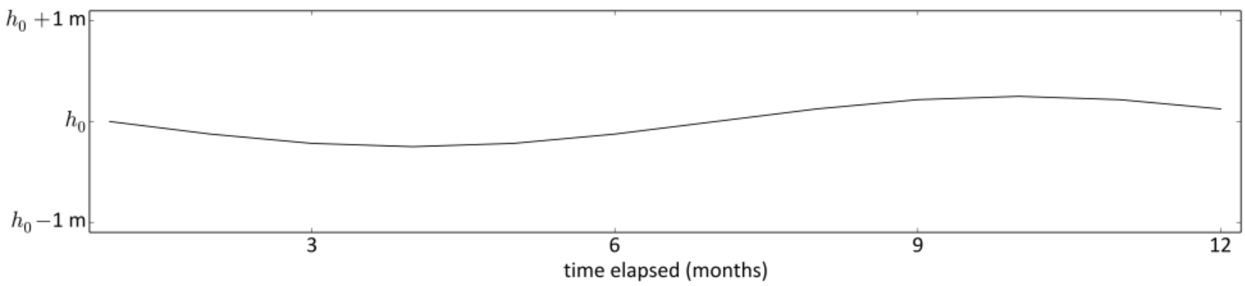

Figure 12. Short term (i.e. seasonal) hypothetical variations in watertable elevation.

When included in the linear sum of component time series, this function features a peak-to-peak amplitude of approximately 0.5 metres, which is consistent with intra-annual climate dynamics.

When the three time series components are combined as a linear sum, the effects of each component may be observed; for example, for a scenario featuring an initial aquifer hydraulic head of 51 m (Figure 13).

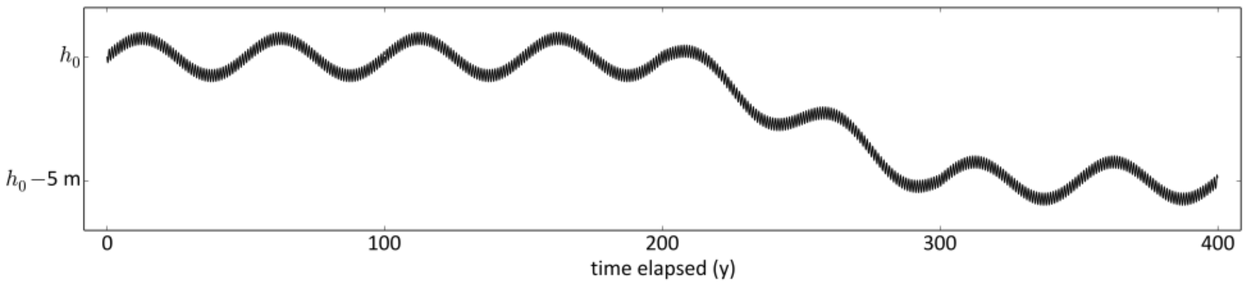

Figure 13. Combined synthetic regional water table variation scenario.

These temporal dynamics are assumed to represent watertable fluctuations along the inflow (i.e. right-hand) boundary of the conceptual model. A hydraulic gradient of 4 m / 5000 m (i.e., 8 x 10^{-4}) is assumed; therefore watertable fluctuations occurring along the outflow (i.e., left-hand) boundary are the same, with an initial value that is 4 m lower than observed on the inflow boundary.

It must be stressed, however, that this water level regime scenario is for illustration purposes only. In practice, it is assumed that the regional scale groundwater flow model would be used to set various boundary conditions for use in a wetland–groundwater interaction model.

See Appendices A.3 and A.4 for example BAS6 and CHD package files and see Appendices B.1 and B.2 for the Python language scripts used to automate the preparation of these input files.

3.4.6 GROUNDWATER STORAGE (d*S*/d*t*)

Fluxes to and from groundwater storage were calculated using the specified storage coefficients, i.e. using a specific yield value of 0.1 for unconfined model cells and 5 x 10^{-3} m^{-1} for confined model cells.

3.4.7 NUMERICAL SOLVER

The discretised 2-D groundwater flow equation, including the wetland water balance equation, was solved using the Preconditioned Conjugate-Gradient solver (Hill, 1990). The maximum number of outer (i.e. MXITER) and inner (i.e. ITER1) solver iterations were set to 50 and 30 respectively. Values for maximum hydraulic head residual (i.e. HCLOSE) and maximum flux residual (i.e. RCLOSE) were specified as 0.01 m and

0.1 m^3.d^{-1} respectively. Other variable values used by the PCG package input file are detailed in Appendix A.7.

3.5 Results of the demonstration model

Each of the numerical solutions of the modelled scenarios featured a cumulative residual error of < 1%. Simulation results are reported as: (1) wetland surface water levels (3.5.1); (2) box and whisker plots of the minimum and maximum annual fractions of wetland inundated (3.5.2); and, (3) wetland salinity risk indices (3.5.3).

3.5.1 WETLAND SURFACE WATER LEVELS

Time series of wetland surface water level were computed for a regional flow-through wetland type featuring surface water additions equivalent to annual water level increases of 0.0, 0.5, 1.0 and 1.5 m (Figure 14).

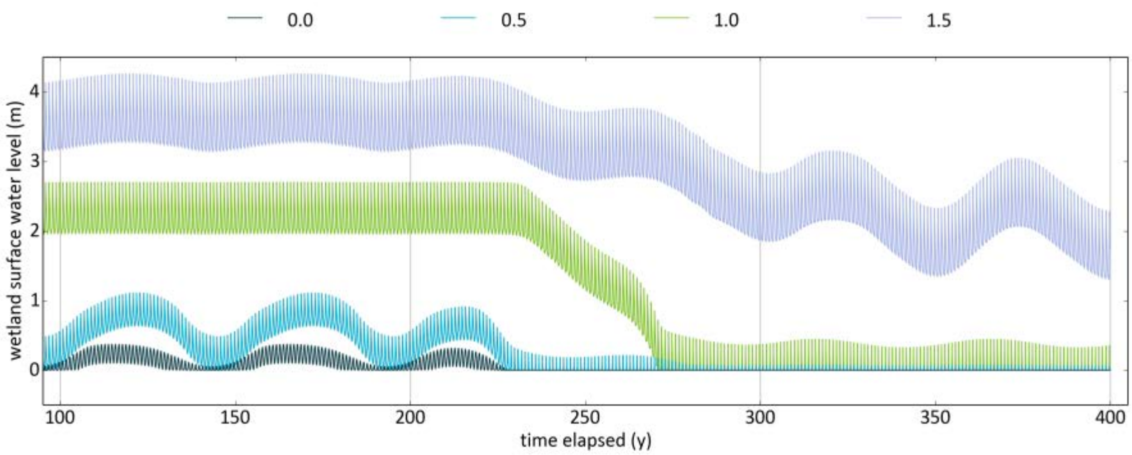

Figure 14. Time series of wetland surface water level for a regional flow-through wetland type with surface water additions equivalent to annual water level increases of 0.0, 0.5, 1.0 and 1.5 m.

Essentially, these simulations suggest that prior to the water table decline (commencing at 200 years elapsed), the wetland would be inundated for the majority of the time without the addition of surface water (black series). Continuous inundation would occur if surface water additions equivalent to annual water level increases of 0.5 m (light blue), 1.0 m (green) or 1.5 m (dark blue) were applied. Conversely, after a period of water table decline (i.e. after 300 years elapsed), of the four surface water addition regimes tested, only an addition equivalent to an annual water level increase of 1.5 m (dark blue) would be sufficient to result in continuous inundation. Surface water addition equivalent to an annual water level increase of 1.0 m (green) would be sufficient to maintain natural wetland dynamics of alternating wetting and drying cycles. Note also that surface water additions equivalent to wetland water level increases of 1.0 m (green) and 1.5 m (light blue) result in water levels in excess of the maximum wetland depth; in practice, this would lead to inundation of areas adjacent to the wetland. Such over-estimation of surface water persistence (and hence, level) may be attributed (at least in part) to the underestimation of wetland infiltration (i.e. $Q_{SW \rightarrow GW}$) by the conductance–based approximation of vadose zone flow used in the present study. Quantification of this potential for overestimation could be considered in future studies, particularly through comparisons to field data.

3.5.2 STATISTICAL SUMMARIES OF WETLAND INUNDATION

Box and whisker plots of minimum and maximum fraction of wetland inundated were computed for (a) the pre-water table decline period and (b) the post-water table decline period for a regional flow-through wetland type with annual surface water additions equivalent to water level increases of 0.0, 0.5, 1.0 and 1.5 m.

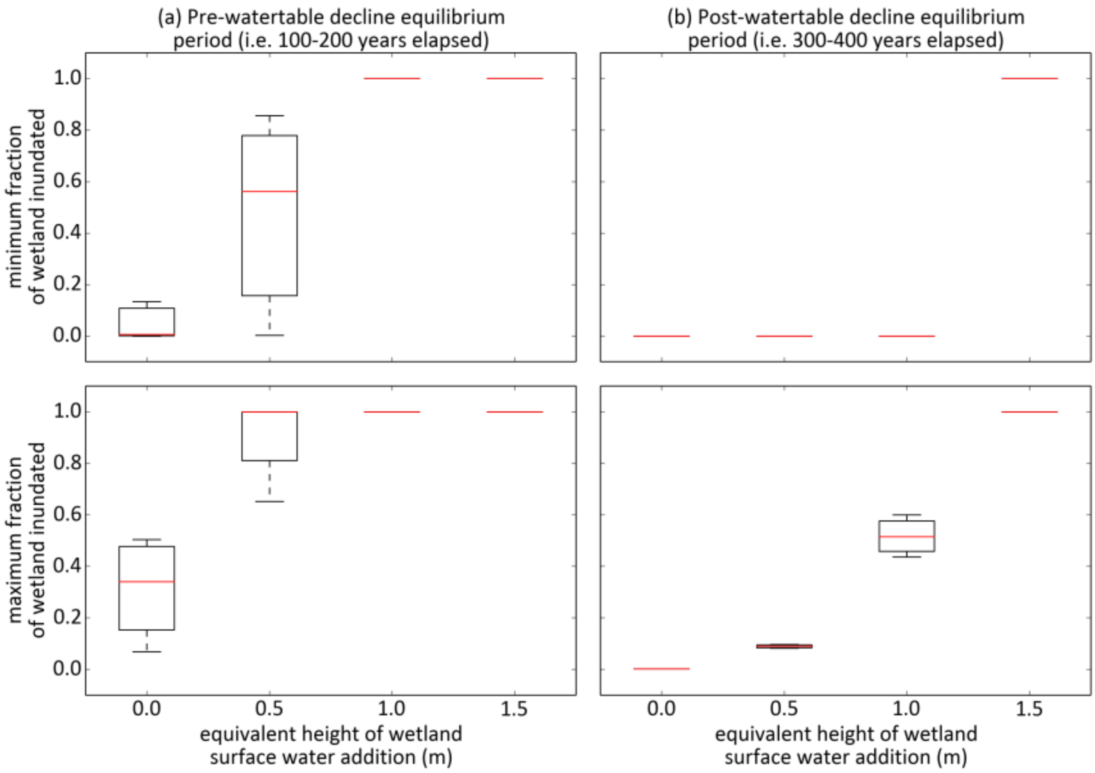

Figure 15. Box and whisker plots of minimum and maximum fraction of wetland inundated for (a) pre-water table decline period and (b) post-water table decline period for a regional flow-through wetland type with annual surface water additions equivalent to water level increases of 0.0, 0.5, 1.0 and 1.5 m. Boxes indicate the interquartile range (IQR) of each data series, whiskers indicate the extent 1.5 times beyond the IQR, and red lines indicate median values.

During the pre-water table decline period (i.e. 100–200 years elapsed), wetlands receiving surface water additions equivalent to wetland water level increases of 1.0 m or 1.5 m remained continuously inundated (i.e. the minimum and maximum fraction of inundation would be equal to unity at all times). The minimum inundated fraction of a wetland receiving zero surface water addition ranged from zero to 0.1, while the maximum inundated fraction varied from 0.1 to 0.5. The minimum inundated fraction of a wetland receiving surface water addition equivalent to a wetland water level increase of 0.5 m ranged from 0.7 to unity, while the maximum inundated fraction varied from zero to 0.8.

During the post-water table decline period (i.e. 300–400 years elapsed), a wetland receiving zero surface water addition remained continuously dry for the duration of the period. Conversely, a wetland receiving surface water addition equivalent to a wetland water level increase of 1.5 m remained continuously inundated for the duration of the period. The inundated fractions of wetlands receiving surface water addition equivalent to wetland water level increases of 0.5 m and 1.0 m ranged from a 0.0–0.1 and 0.0–0.5 respectively.

3.5.3 WETLAND SALINISATION RISK METRIC

A simple wetland salinisation risk metric was computed for a regional flow-through wetland with annual surface water additions equivalent to water level increases of 0.0, 0.5, 1.0 and 1.5 m. High metric values (i.e., ≃1.0) indicate a high likelihood of wetland salinisation occurring.

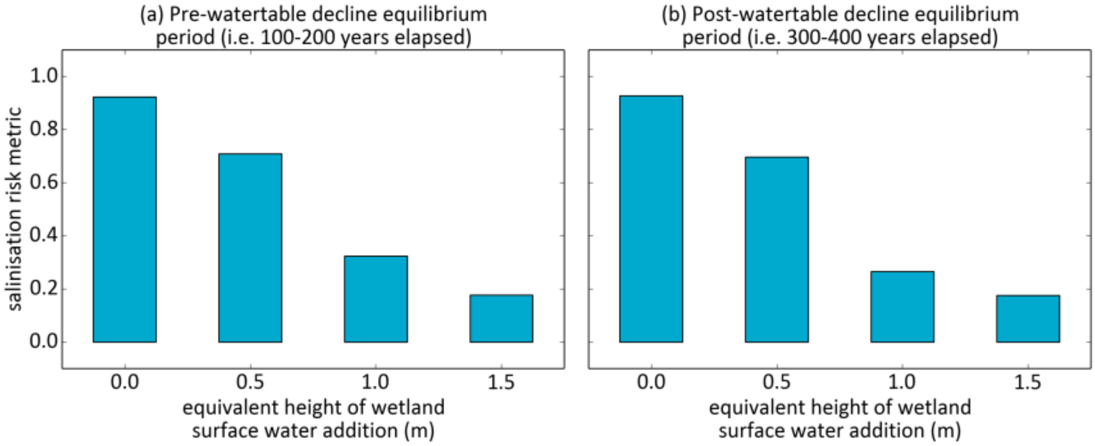

Figure 16. Wetland salinisation risk metrics for a regional flow-through wetland with annual surface water additions equivalent to water level increases of 0.0, 0.5, 1.0 and 1.5 m. High metric values (i.e., ≃1.0) indicate a high likelihood of wetland salinisation occurring.

Salinisation risk metric results are consistent with expectation; i.e., wetlands receiving zero surface water inflows are associated with high salinisation risk while increases in the magnitude of wetland surface water inflow correlate with reductions in salinisation risk. These results highlight the flushing effect of surface water inflows, which act to remove evapo-concentrated salts stored at the surface and in the vadose zone of wetlands.

4 Linking the wetland–groundwater interaction modelling approach with the regional model

At the time of reporting, the development of the Lower Limestone Coast Prescribed Wells Area (LLC PWA) regional scale groundwater flow model was not at a stage whereby a means of linking a local scale wetland–groundwater interaction model could be trialled and tested. Instead, a potential method by which such linkage could be undertaken is now discussed.

One of the most important aspects when setting up a wetland–groundwater interaction model is to ensure that the boundary conditions used are appropriate for the type of wetland and the regional context. The LLC PWA region is large and spatially variable. It features a north–south climate gradient, a spatially variable depth to bedrock (which also varies along a north–south gradient (Harrington and Lamontagne, 2013), and a complex topography (that is, a significant potential for the development of local flow systems). Here it is proposed to define 'hydrogeological' subregions and to define representative boundary conditions for the wetland model at this scale (that is, time series of sub-regional water table variation regimes).

Hydrogeological sub-regions could be defined based on (1) the presence of upward vertical flow due to shallow bedrock intrusion and (2) whether wetlands are located in regional recharge or discharge areas. As a part of the development of the regional model, a review of the hydrostratigraphy of the LLC PWA was undertaken (Harrington and Lamontagne, 2013). The northern half of the LLC PWA generally features a shallower depth to bedrock and a generally drier climate in comparison to the southern area (Figure 17). These features have resulted in the development of numerous interdunal and local scale groundwater flow systems. On this basis, northern and southern hydrogeological sub-regions could be defined, which would also capture, to some extent, effects of the latitudinal climatic gradient present in the study area.

(a)

(b)

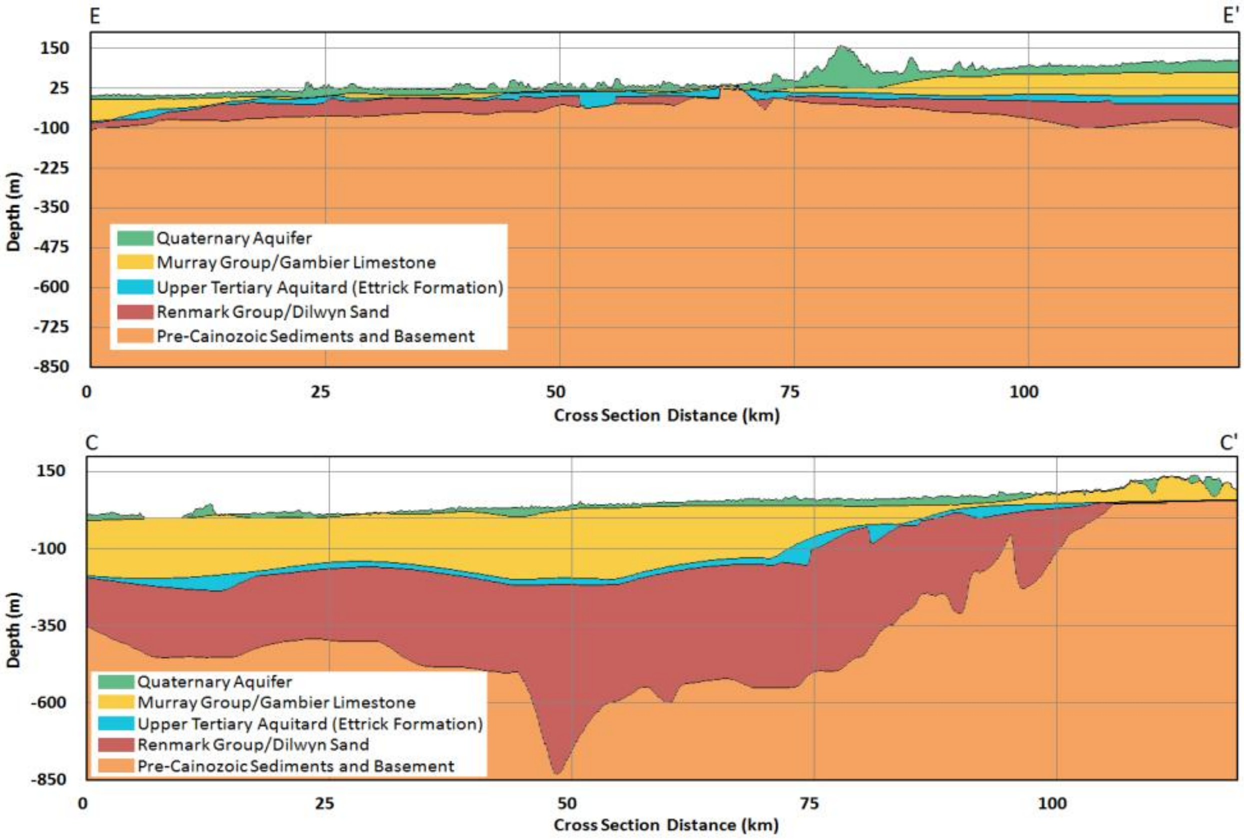

Figure 17. Geological cross-sections for (a) the northern and (b) the southern LLC PWA. For exact locations of these transects, see Harrington and Lamontagne (2013).

In a subsequent step, the regional scale model could be used to broadly define recharge and discharge areas in the study area. This would be desirable because, in the LLC PWA, water tables in regional recharge zones typically feature greater temporal variation than those located in regional discharge areas (Taylor et al., 2015). Specifically, vertical velocity vector outputs from the steady state regional scale model (e.g., Figure 18) could be used in a first-order approach to identifying recharge and discharge regions. In the southern area (where bedrock intrusion is limited), the spatial distribution of recharge and discharge zones suggests a single continuous groundwater flow system. A large recharge zone is located up-gradient (i.e., in the east) and a large discharge zone is located along the coastline. Conversely, in the northern area, where bedrock control is thought to occur, the spatial distribution of recharge and discharge zones is more complex (Figure 18). In particular, one large discharge zone is located inland and corresponds to the location of the Bool Lagoon Boinka (Taylor et al., 2015). Thus the LLC PWA contains at least four 'hydrogeological' subregions:

1) regional recharge areas over shallow bedrock;
2) regional recharge areas over deep bedrock;
3) regional discharge areas over shallow bedrock; and,
4) regional discharge areas over deep bedrock.

In practice, it may be necessary to subdivide the LLC PWA further; for example, discharge areas along the coast may feature different water table regimes to those of discharge areas located inland. Once sub-regions are defined, the regional model could be interrogated to extract a representative water table variation regime from each sub-region. How this could be undertaken would require further evaluation once development of the regional scale groundwater flow model is completed.

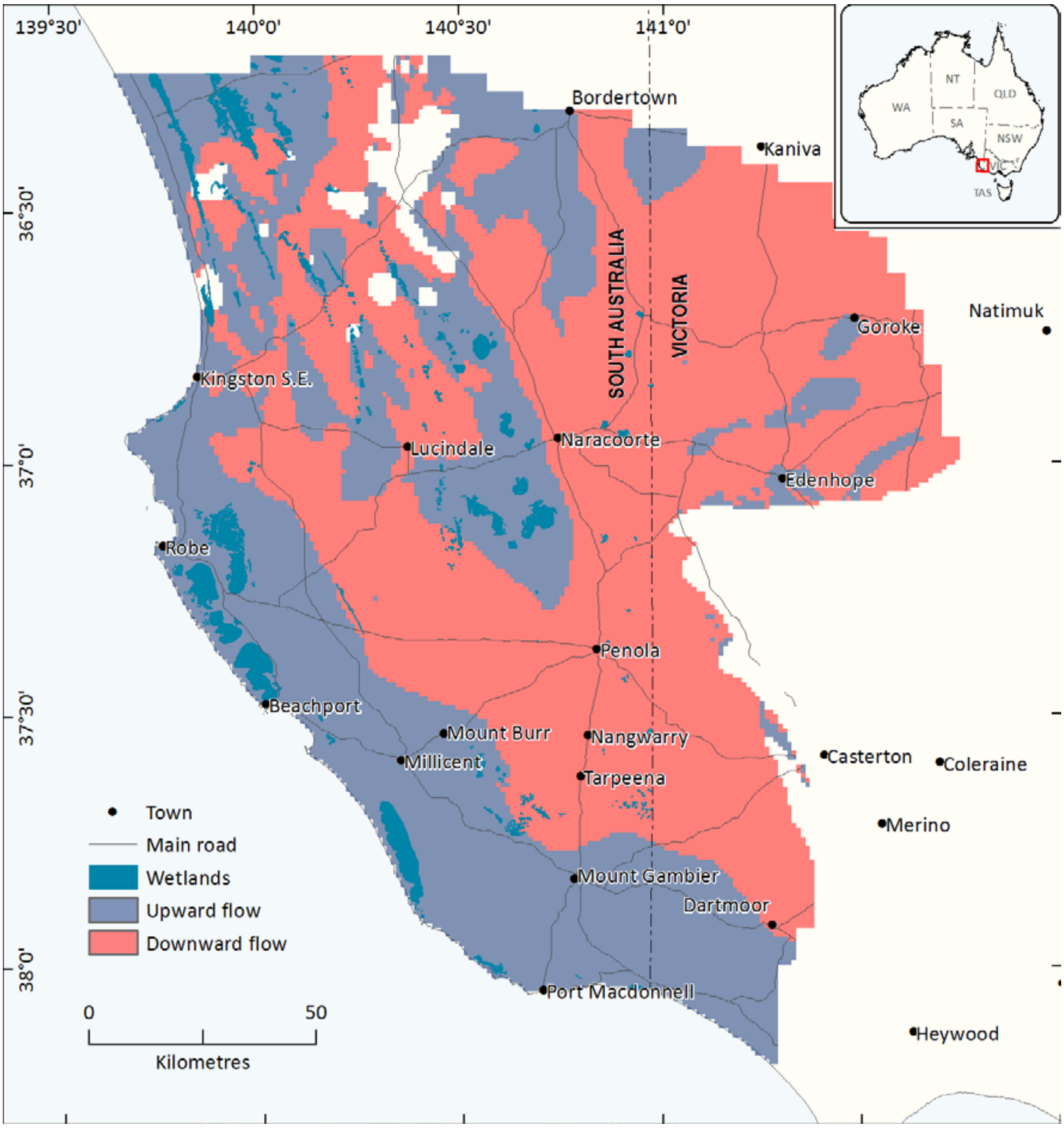

Figure 18. Estimated areas of upward (purple) and downward (red) groundwater flow, based on the vertical velocity vectors derived from the steady state regional scale groundwater flow model (L. Morgan, unpublished data).

5 Conclusions

An approach to simulating wetland–groundwater interactions using the saturated groundwater flow code MODFLOW was described but has not been 'road tested' as yet. In particular, whilst a proposed strategy to link the regional scale groundwater flow model with local scale wetland–groundwater interaction models was proposed, this strategy could not be evaluated because the regional model had not been completed at the time of writing this report. This will need to be addressed as a part of future model developments.

The proposed approach to wetland–groundwater interaction modelling is generic in nature and should not be used for making specific predictions for particular Lower Limestone Coast Prescribed Wells Area (LLC PWA) wetlands. The aim of the approach is to characterise the behaviour of a population of wetlands (particularly shallow wetlands such as deflation basins) at a sub-regional scale in the study area. This approach should be sufficient to evaluate whether the surface water regime of given wetlands will become generally wetter or drier at the sub-regional scale under different scenarios of change in climate, land-use and water allocation policy. For example, in the synthetic dataset demonstration, the temporal dynamics of wetlands featuring various surface water augmentation regimes were found to change from perennial to ephemeral or episodic in response to lowering of the regional water table (i.e., reflecting reductions in regional recharge). Further analysis could involve evaluation of the sensitivity of wetland responses to local factors, such as supplementation with surface water, the presence or absence of a clogging layer, or variations in wetland morphometry.

Application of the modelling approach to a synthetic dataset indicated that the duration of inundation is positively correlated with the magnitude of surface water addition. While this result is intuitive, the utility of the wetland model lies in the ability to identify potential thresholds or 'tipping points': for example, under a future scenario featuring declining regional recharge (and therefore water tables), the conditions under which changes in wetlands water level regime (e.g. from perennial to episodic) may occur could be identified. The sensitivity of such thresholds to the use of managed surface water supplementation could also be assessed.

Impediments to the creation of wetland–groundwater interaction models for <u>specific</u> LLC PWA wetlands remain numerous at present. These include:

1) The monitoring of hydrological data is currently sparse (in both space and time) for practically all wetlands in the region. Without the availability of long-term monitoring datasets, it is not possible to calibrate hydrological models to a level where predictions can be made with a reasonable degree of certainty.
2) Water balances for most LLC PWA wetlands are complex. Evaporation (*E*) and evapotranspiration (*ET*) fluxes – especially in regional groundwater discharge areas – are the largest components of the water balances for the region's wetlands. However, the approaches used to estimate *E* and *ET* in the present study are simple and based on regional meteorological data. *E* and *ET* in LLC PWA wetlands should be independently measured and compared to the approach used here to ensure that these are representative.
3) For most LLC PWA wetlands, the local sedimentary environment is not well-characterised. In particular, the presence, extent and hydraulic properties of wetland clogging layers must be known in order to evaluate whether variably-saturated (i.e. 'perched') conditions can occur below a wetland if the regional water table drops below the elevation of the wetland bed.

Subsequent steps in the development and application of the wetland–groundwater interaction model could include:

1) Further refinement of the proposed down-scaling strategy to define boundary conditions from the regional model;

2) Quantification of the potential for overestimation of surface water persistence resulting from the conductance-based approximations of vadose zone flow;
3) A scenario analysis (‘road testing’) for a selection of potential long-term changes in land-use, climate or water allocation policy by combining both the regional groundwater flow and wetland–groundwater interaction models, with the aim to evaluate whether the proposed downscaling strategy is practicable; and,
4) Undertake a remote sensing study to characterise the historical water level regime of LLC PWA wetlands from 1970 to the present day. This could be based on a refinement of the approach recently proposed by Deane et al. (2015) to evaluate the inundation regime of South-East wetlands. Once available, this historical water level regime database could be used to further develop and refine the wetland–groundwater interaction model.

Lastly, neither the regional model of Morgan et al. (2015) nor the wetland–groundwater interaction model proposed here include solute transport and thus cannot be used to evaluate variations in wetland salinity over time, other than through the salinity risk index proposed here. Adding solute transport to these models is feasible and could be considered as a part of future model development. However, the development of wetland salt mass balance models would require that long-term monitoring of salinity (i.e., electrical conductivity) be undertaken. Currently, while the availability of wetland surface water level data is quite low, the monitoring of wetland salinity data is even less common.

Appendix A Example MODFLOW-OWHM input files

The MODFLOW–OWHM input files used to create the synthetic wetland–groundwater interaction demonstration model described in Section 3 are presented here (in part) for demonstration purposes.

A.1 Discretisation (DIS) package

```
    6   1  50   4800   4   2
    0   0   0      0   0   0
CONSTANT 100.0
CONSTANT 1.0
INTERNAL 1 (10F8.2) 1
   60.00   60.08   60.16   60.24   60.33   60.41   60.49   60.57   60.65   60.73
   60.82   60.90   60.98   61.06   61.14   61.22   61.31   61.39   61.47   61.55
   61.63   61.71   61.80   61.88   61.96   62.04   62.12   62.20   62.29   62.37
   62.45   62.53   62.61   62.69   62.78   62.86   62.94   63.02   63.10   63.18
   63.27   63.35   63.43   63.51   63.59   63.67   63.76   63.84   63.92   64.00
INTERNAL 1 (10F8.2) 1
   50.00   50.08   50.16   50.24   50.33   50.41   50.49   50.57   50.65   50.73
   50.82   50.90   50.98   51.06   51.14   51.22   51.31   51.39   51.47   51.55
   59.63   59.71   59.80   59.88   59.96   60.04   60.12   60.20   60.29   60.37
   52.45   52.53   52.61   52.69   52.78   52.86   52.94   53.02   53.10   53.18
   53.27   53.35   53.43   53.51   53.59   53.67   53.76   53.84   53.92   54.00
INTERNAL 1 (10F8.2) 1
   40.00   40.08   40.16   40.24   40.33   40.41   40.49   40.57   40.65   40.73
   40.82   40.90   40.98   41.06   41.14   41.22   41.31   41.39   41.47   41.55
   41.63   41.71   41.80   41.88   41.96   42.04   42.12   42.20   42.29   42.37
   42.45   42.53   42.61   42.69   42.78   42.86   42.94   43.02   43.10   43.18
   43.27   43.35   43.43   43.51   43.59   43.67   43.76   43.84   43.92   44.00
CONSTANT 30.0
CONSTANT 20.0
CONSTANT 10.0
CONSTANT  0.0
  30.0    30   1.0     TR
...
```

A.2 Layer-Property Flow (LPF) package

```
    50  -1e30      0
     3      3      3      3      3     3
     0      0      0      0      0     0
   1.0    1.0    1.0    1.0    1.0   1.0
     0      0      0      0      0     0
     0      0      0      0      0     0
CONSTANT   10.0
CONSTANT   10.0
CONSTANT   5e-3
CONSTANT   0.1
CONSTANT   10.0
CONSTANT   10.0
CONSTANT   5e-3
CONSTANT   0.1
CONSTANT   10.0
CONSTANT   10.0
CONSTANT   5e-3
CONSTANT   0.1
CONSTANT   10.0
CONSTANT   10.0
CONSTANT   5e-3
CONSTANT   0.1
CONSTANT   10.0
CONSTANT   10.0
CONSTANT   5e-3
CONSTANT   0.1
CONSTANT   10.0
CONSTANT   10.0
CONSTANT   5e-3
CONSTANT   0.1
```

A.3 Basic (BAS6) package

```
FREE
INTERNAL 1 (10I4) 1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   0   0   0   0   0   0   0   0   0   0
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
INTERNAL 1 (10I4) 1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
INTERNAL 1 (10I4) 1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
INTERNAL 1 (10I4) 1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
INTERNAL 1 (10I4) 1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
INTERNAL 1 (10I4) 1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
   1   1   1   1   1   1   1   1   1   1
-9999.0
INTERNAL 1 (10F8.2) 1
   58.00   58.08   58.16   58.24   58.33   58.41   58.49   58.57   58.65   58.73
   58.82   58.90   58.98   59.06   59.14   59.22   59.31   59.39   59.47   59.55
   59.63   59.71   59.80   59.88   59.96   60.04   60.12   60.20   60.29   60.37
   60.45   60.53   60.61   60.69   60.78   60.86   60.94   61.02   61.10   61.18
   61.27   61.35   61.43   61.51   61.59   61.67   61.76   61.84   61.92   62.00
INTERNAL 1 (10F8.2) 1
   58.00   58.08   58.16   58.24   58.33   58.41   58.49   58.57   58.65   58.73
   58.82   58.90   58.98   59.06   59.14   59.22   59.31   59.39   59.47   59.55
   59.63   59.71   59.80   59.88   59.96   60.04   60.12   60.20   60.29   60.37
   60.45   60.53   60.61   60.69   60.78   60.86   60.94   61.02   61.10   61.18
   61.27   61.35   61.43   61.51   61.59   61.67   61.76   61.84   61.92   62.00
INTERNAL 1 (10F8.2) 1
   58.00   58.08   58.16   58.24   58.33   58.41   58.49   58.57   58.65   58.73
   58.82   58.90   58.98   59.06   59.14   59.22   59.31   59.39   59.47   59.55
   59.63   59.71   59.80   59.88   59.96   60.04   60.12   60.20   60.29   60.37
   60.45   60.53   60.61   60.69   60.78   60.86   60.94   61.02   61.10   61.18
   61.27   61.35   61.43   61.51   61.59   61.67   61.76   61.84   61.92   62.00
...
```

A.4 Time-variant specified-head(CHD) package

```
        10
        10
         2         1         1    58.000    58.000
         3         1         1    58.000    58.000
         4         1         1    58.000    58.000
         5         1         1    58.000    58.000
         6         1         1    58.000    58.000
         2         1        50    62.000    62.000
         3         1        50    62.000    62.000
         4         1        50    62.000    62.000
         5         1        50    62.000    62.000
         6         1        50    62.000    62.000
        10
         2         1         1    57.883    57.883
         3         1         1    57.883    57.883
         4         1         1    57.883    57.883
         5         1         1    57.883    57.883
         6         1         1    57.883    57.883
         2         1        50    61.883    61.883
         3         1        50    61.883    61.883
         4         1        50    61.883    61.883
         5         1        50    61.883    61.883
         6         1        50    61.883    61.883
        10
         2         1         1    57.799    57.799
         3         1         1    57.799    57.799
         4         1         1    57.799    57.799
         5         1         1    57.799    57.799
         6         1         1    57.799    57.799
         2         1        50    61.799    61.799
         3         1        50    61.799    61.799
         4         1        50    61.799    61.799
         5         1        50    61.799    61.799
         6         1        50    61.799    61.799
        10
         2         1         1    57.774    57.774
         3         1         1    57.774    57.774
         4         1         1    57.774    57.774
         5         1         1    57.774    57.774
         6         1         1    57.774    57.774
         2         1        50    61.774    61.774
         3         1        50    61.774    61.774
         4         1        50    61.774    61.774
         5         1        50    61.774    61.774
         6         1        50    61.774    61.774
        10
         2         1         1    57.815    57.815
         3         1         1    57.815    57.815
         4         1         1    57.815    57.815
         5         1         1    57.815    57.815
         6         1         1    57.815    57.815
         2         1        50    61.815    61.815
         3         1        50    61.815    61.815
         4         1        50    61.815    61.815
         5         1        50    61.815    61.815
         6         1        50    61.815    61.815
...
```

A.5 Riparian Evapotranspiration (RIP-ET) package

```
    50  1  -1  -1
   2  17
    land       0.0      10.0     9.190418E-04   9.190418E-04      17
   1.000000E-02   3.900000E-01   1.000000E-01   5.000000E-02   5.000000E-02
   3.184581E-01   0.000000E+00  -4.111058E-03  -6.107007E-03  -1.295877E-02  -
    wetland    0.0      10.0     9.190418E-04   0.000000E+00      17
   1.000000E-02   3.900000E-01   1.000000E-01   5.000000E-02   5.000000E-02
   3.184581E-01   0.000000E+00  -4.111058E-03  -6.107007E-03  -1.295877E-02  -   50
   2   1   1   1
   60.00     1.0     0.0
   2   1   2   1
   60.08     1.0     0.0
   2   1   3   1
   60.16     1.0     0.0
...
-1
-1
-1
-1
-1
...
```

A.6 Lake (LAK) package

```
        10        50
       0.0       100       0.0
     59.63     59.63     61.63
     59.71     59.71     61.71
     59.80     59.80     61.80
     59.88     59.88     61.88
     59.96     59.96     61.96
     60.04     60.04     62.04
     60.12     60.12     62.12
     60.20     60.20     62.20
     60.29     60.29     62.29
     60.37     60.37     62.37
         1         1         0
INTERNAL 1 (10I4) 1
   0   0   0   0   0   0   0   0   0   0
   0   0   0   0   0   0   0   0   0   0
   1   2   3   4   5   6   7   8   9  10
   0   0   0   0   0   0   0   0   0   0
   0   0   0   0   0   0   0   0   0   0
CONSTANT 0
CONSTANT 0
CONSTANT 0
CONSTANT 0
CONSTANT 0
CONSTANT 0.227
CONSTANT 0.0
CONSTANT 0.0
CONSTANT 0.0
CONSTANT 0.0
CONSTANT 0.0
         0
    0.0007    0.0054    0.0000    0.0000
    0.0007    0.0054    0.0000    0.0000
    0.0007    0.0054    0.0000    0.0000
    0.0007    0.0054    0.0000    0.0000
    0.0007    0.0054    0.0000    0.0000
    0.0007    0.0054    0.0000    0.0000
    0.0007    0.0054    0.0000    0.0000
    0.0007    0.0054    0.0000    0.0000
    0.0007    0.0054    0.0000    0.0000
    0.0007    0.0054    0.0000    0.0000
...
```

A.7 Preconditioned Conjugate-Gradient solver (PCG) package

```
   50     30      1
  1e-2   1e-1    1.0      2      0      3    1.0
```

A.8 Output Control (OC) package

```
HEAD PRINT FORMAT 0
HEAD SAVE UNIT 30
PERIOD 1 STEP 30
 PRINT HEAD
 PRINT BUDGET
 SAVE HEAD
 SAVE BUDGET
PERIOD 2 STEP 30
 PRINT HEAD
 PRINT BUDGET
 SAVE HEAD
 SAVE BUDGET
PERIOD 3 STEP 30
 PRINT HEAD
 PRINT BUDGET
 SAVE HEAD
 SAVE BUDGET
...
```

A.9 Gauge (GAG) package

```
  10
  -1   51
  -2   52
  -3   53
  -4   54
  -5   55
  -6   56
  -7   57
  -8   58
  -9   59
  -10  60
```

Appendix B Scripts used to write model input files

The Python language scripts used to create the MODFLOW–OWHM input files for the synthetic wetland–groundwater interaction demonstration model described in Section 3 are presented here for demonstration purposes.

B.1 Write BAS6 package

```
import numpy as np
from math import pi, sin

ih = 58.0

bas = open( 'bas.dat', 'w' )

bas.write( 'FREE\n' )
bas.write( 'INTERNAL 1 (10I4) 1\n' )
bas.write( '   1   1   1   1   1   1   1   1   1   1\n' )
bas.write( '   1   1   1   1   1   1   1   1   1   1\n' )
bas.write( '   0   0   0   0   0   0   0   0   0   0\n' )
bas.write( '   1   1   1   1   1   1   1   1   1   1\n' )
bas.write( '   1   1   1   1   1   1   1   1   1   1\n' )

for i in range( 0, 5 ):
    bas.write( 'INTERNAL 1 (10I4) 1\n' )
    bas.write( '   1   1   1   1   1   1   1   1   1   1\n' )
    bas.write( '   1   1   1   1   1   1   1   1   1   1\n' )
    bas.write( '   1   1   1   1   1   1   1   1   1   1\n' )
    bas.write( '   1   1   1   1   1   1   1   1   1   1\n' )
    bas.write( '   1   1   1   1   1   1   1   1   1   1\n' )

bas.write( '-9999.0\n' )

IH = np.reshape( np.linspace( ih, ih+4.0, 50 ), [ 5, 10 ] )

for i in range( 0, 6 ):
    bas.write( 'INTERNAL 1 (10F8.2) 1\n' )
    for j in range( 0, 5 ):
        for k in range( 0, 10 ):
            bas.write( '%8.2f'% IH[ j, k ] )
        bas.write( '\n' )

bas.close()
```

B.2 Write CHD package

```
from math import pi, sin

chd = open( 'chd.dat', 'w' )

nsp = 4800
ih = 58.0
n = 0
h = ih

chd.write( '        10\n' )

for sp in range( 0, nsp ):
    chd.write( '        10\n' )
    if sp < 2401:
        A = ih
    elif sp > 2401 and sp < 3600:
        A = A - ( 5. / 1200. )
    B = 0.75 * sin ( ( sp / 12. ) * ( 2. * pi / 50. ) )
    C = 0.25 * sin ( 2 * pi * ( ( sp / 12. ) + ( 6. / 12. ) ) )
    h = A + B + C
    for lay in range( 2, 7, 1 ):
        chd.write( '%10i%10i%10i%10.3f%10.3f%1s'% ( lay, 1, 1, h, h, '\n' ) )
    for lay in range( 2, 7, 1 ):
        chd.write( '%10i%10i%10i%10.3f%10.3f%1s'% ( lay, 1, 50, h+4.0, h+4.0, '\n' ) )

chd.close()
```

B.3 Write RIP-ET package

```
import numpy as np

# subroutine for counting number of lines in input file
def file_len(fname):
    with open(fname) as f:
        for i, l in enumerate(f):
            pass
    return i + 1

# get input data: z and r
z, r = [], []
fn = '1132.csv'
flen = file_len( fn )
f = open( fn, 'r' )
for i in range( 0, flen ):
    l = f.readline()
    if i > 0:
        l = l.strip()
        l = l.split( ',' )
        z.append( float( l[0] ) )
        r.append( float( l[1] ) )
f.close()
z = np.array( z )
r = np.array( r ) * -1.

# convert r from mm/y to m/d
r = r / 1000.0 / 365.25

# insert two dummy elevations and values below input data series
z = np.insert( z, len(z), z[-1]+3.9 )
z = np.insert( z, len(z), z[-2]+4.0 )
r = np.insert( r, len(r), r[-1] )
r = np.insert( r, len(r), 0.0 )

# compute fdh and fdR
N = len(z)-1
fdh = np.zeros( [ N ] )
fdR = np.zeros( [ N ] )
for i in range( 0, N ):
    # fdh is change in depth as fraction of total depth
    fdh[i] = ( z[i+1] - z[i] ) / ( z[-1] - z[0] )
    # fdR is change in flux as fraction of minimum flux
    fdR[i] = ( r[i+1] - r[i] ) / abs( r.max() )

# flip vectors as RIP input is read from bottom to top
fdh = fdh[::-1]
fdR = fdR[::-1]

# create separate fdR arrays for land and wetland areas
# (only top [i.e. last] value is different)
fdR_ln = fdR.copy()
fdR_ln[ N-1 ] = 0.0
fdR_wl = fdR.copy()

# read land surface elevations from DIS file
Z = []
f = open( 'dis.dat', 'r' )
for i in range( 0, 5):
    l = f.readline()
for i in range( 0, 5):
    l = f.readline()
    l = l.strip()
    l = l.split()
    for item in l:
        Z.append( float( item ) )
f.close()
for i in range( 20, 30 ):
```

```
    Z[i] = Z[i]-2.0

# create function indicator arrays
L = np.ones( [ 50 ] )
for i in range( 20, 30 ):
    L[i] = 0.0
W = np.zeros( [ 50 ] )
for i in range( 20, 30 ):
    W[i] = 1.0

# write RIP file
f = open( 'rip.dat', 'w' )
f.write( '   50  1  -1  -1\n' )
f.write( '   2' )
f.write( '%4i%s'% ( N, '\n' ) )

f.write( '    land   ' )
f.write( '%7.1f%10.1f%17.6e'% ( z.min(), z[-1] - z[0], abs( r.max() ) ) )
f.write( '%15.6e%8i%s'% ( abs( r.max() ), N, '\n' ) )
for i in range( 0, N ):
    f.write( '%15.6e'% fdh[i] )
f.write( '\n' )
for i in range( 0, N ):
    f.write( '%15.6e'% fdR_ln[i] )
f.write( '\n' )

f.write( '    wetland' )
f.write( '%7.1f%10.1f%17.6e'% ( z.min(), z[-1] - z[0], abs( r.max() ) ) )
f.write( '%15.6e%8i%s'% ( 0.0, N, '\n' ) )
for i in range( 0, N ):
    f.write( '%15.6e'% fdh[i] )
f.write( '\n' )
for i in range( 0, N ):
    f.write( '%15.6e'% fdR_wl[i] )
f.write( '\n' )

f.write( '   50\n' )
for i in range( 0, 50 ):
    f.write( '%4i%4i%4i%4i%s'% ( 2, 1, i+1, 1, '\n' ) )
    f.write( '%8.2f%8.1f%8.1f%s'% ( Z[i], L[i], W[i], '\n' ) )
for i in range( 0, 4800 ):
    f.write( '-1\n' )
f.close()
```

B.4 Write LAK package

```
from math import pi, sin

nyr = 400
pmx, pcx, pmy, pcy = -1.7, 1.5, -30.,  50.
emx, ecx, emy, ecy = -1.8, 1.3, 100., 140.
dstage = 1.5

lak = open( 'lak.dat', 'w' )

lak.write( '        10        50\n' )
lak.write( '       0.0       100       0.0\n' )
lak.write( '     53.63     51.63     61.63\n' )
lak.write( '     53.71     51.71     61.71\n' )
lak.write( '     53.80     51.80     61.80\n' )
lak.write( '     53.88     51.88     61.88\n' )
lak.write( '     53.96     51.96     61.96\n' )
lak.write( '     54.04     52.04     62.04\n' )
lak.write( '     54.12     52.12     62.12\n' )
lak.write( '     54.20     52.20     62.20\n' )
lak.write( '     54.29     52.29     62.29\n' )
lak.write( '     54.37     52.37     62.37\n' )

for i in range( 0, nyr ):
    for m in range( 0, 12 ):
        px = pi * ( m / 11. ) * pmx + pcx
        py = ( sin( px ) * pmy + pcy ) / ( 365.25 / 12. ) / 1000.
        ex = pi * ( m / 11. ) * emx + ecx
        ey = ( sin( ex ) * emy + ecy ) / ( 365.25 / 12. ) / 1000. * 0.97

        lak.write( '         1         1         0\n' )
        lak.write( 'INTERNAL 1 (10I4) 1\n' )
        lak.write( '   0   0   0   0   0   0   0   0   0   0\n' )
        lak.write( '   0   0   0   0   0   0   0   0   0   0\n' )
        lak.write( '   1   2   3   4   5   6   7   8   9  10\n' )
        lak.write( '   0   0   0   0   0   0   0   0   0   0\n' )
        lak.write( '   0   0   0   0   0   0   0   0   0   0\n' )
        lak.write( 'CONSTANT 0\n' )
        lak.write( 'CONSTANT 0\n' )
        lak.write( 'CONSTANT 0\n' )
        lak.write( 'CONSTANT 0\n' )
        lak.write( 'CONSTANT 0\n' )
        lak.write( 'CONSTANT 0.227\n' )
        lak.write( 'CONSTANT 0.0\n' )
        lak.write( 'CONSTANT 0.0\n' )
        lak.write( 'CONSTANT 0.0\n' )
        lak.write( 'CONSTANT 0.0\n' )
        lak.write( 'CONSTANT 0.0\n' )
        lak.write( '         0\n' )
        for c in range ( 0, 10 ):
            lak.write( '%10.4f'% py )
            lak.write( '%10.4f'% ey )
            lak.write( '%10.4f'% 0. )
            if m == 8:
                lak.write( '%10.4f'% ( -dstage * 100. / 30. ) )
            else:
                lak.write( '%10.4f'% 0. )
            lak.write( '\n' )


lak.close()
```

Appendix C Scripts used to post-process model output files

The Python language scripts used to post-process outputs from the synthetic wetland–groundwater interaction demonstration model described in Section 3 are presented here for demonstration purposes.

C.1 Plot wetland surface water levels versus time

```
import numpy as np
import matplotlib as mpl
import matplotlib.pyplot as pyp
from os import getcwd, chdir, listdir

listdir( getcwd() )

def get_timeseries( direct ):
    print direct
    chdir( direct )
    H = []
    I = open( 'gage01.out', 'r' )
    for i in range( 0, 2 ):
        l = I.readline()
    for i in range( 0, 144000 ):
        l = I.readline()
        l = l.strip()
        l = l.split()
        H.append( float( l[1] ) - 59.63 )
    I.close()
    chdir( '..' )
    return H

maxt, tincr, xtra = 144000, 18000, 1800

green   = [ 120.0 / 255.0, 190.0 / 255.0,   32.0 / 255.0 ]
lt_blue = [   0.0 / 255.0, 169.0 / 255.0 , 206.0 / 255.0 ]
dk_blue = [   0.0 / 255.0,  49.0 / 255.0 ,  60.0 / 255.0 ]
purple  = [ 159.0 / 255.0, 174.0 / 255.0 , 229.0 / 255.0 ]

mpl.rcParams[ 'font.sans-serif' ] = 'Calibri'
mpl.rcParams[ 'font.size' ] = 20

f = pyp.figure( figsize=[ 16, 6 ] )
s = f.add_subplot( 111 )

p = s.plot( [  36000,  36000 ], [ -2, 20 ], '-', c='Gray', alpha=0.5 )
p = s.plot( [  72000,  72000 ], [ -2, 20 ], '-', c='Gray', alpha=0.5 )
p = s.plot( [ 108000, 108000 ], [ -2, 20 ], '-', c='Gray', alpha=0.5 )
p = s.plot( [ 144000, 144000 ], [ -2, 20 ], '-', c='Gray', alpha=0.5 )

t = np.arange( 0, maxt )

p15 = s.plot( t, get_timeseries( '5p2d_SW1.5_FLT' ), '-', c=purple,  label='1.5' )
p10 = s.plot( t, get_timeseries( '5p2c_SW1.0_FLT' ), '-', c=green,   label='1.0' )
p05 = s.plot( t, get_timeseries( '5p2b_SW0.5_FLT' ), '-', c=lt_blue, label='0.5' )
p00 = s.plot( t, get_timeseries( '5p2a_SW0.0_FLT' ), '-', c=dk_blue, label='0.0' )

s.set_xlabel( 'time elapsed (y)' )
s.set_xticks( np.arange( 0, maxt+tincr, tincr ) )
s.set_xticklabels( np.arange( 0, 450, 50 ) )
s.set_xlim( 36000-xtra, maxt+xtra )
```

```
s.set_ylabel( 'wetland surface water level (m)' )
s.set_yticks( np.arange( 0, 5, 1 ) )
s.set_ylim( -0.5, 4.5 )
handles, labels = s.get_legend_handles_labels()
l = s.legend( handles[::-1], labels[::-1], loc=9, ncol=5, bbox_to_anchor=[ 0.5, 1.2 ],
columnspacing=5 )
pyp.setp( l.get_texts(), fontsize=20 )
l.get_frame().set_edgecolor( 'white' )

pyp.tight_layout()
pyp.savefig( 'Plot_wetland_level_vs_time.png', dpi=200, bbox_extra_artists=(l,),
bbox_inches='tight' )
pyp.close( f )
```

C.2 Plot statistical summaries of wetland surface water level variation

```
import numpy as np
import matplotlib as mpl
import matplotlib.pyplot as pyp


maxt, tincr, xtra = 144000, 18000, 1800
zbot01, zbot10 = 59.63, 60.37

INfdr = [ '5p2a_SW0.0_FLT/',
          '5p2b_SW0.5_FLT/',
          '5p2c_SW1.0_FLT/',
          '5p2d_SW1.5_FLT/'  ]

data = np.zeros( [ len( INfdr ), 144000 ] )
for j in range( 0, len( INfdr ) ):
     H = []
     INfnm = INfdr[j] + 'gage01.out'
     INf = open( INfnm, 'r' )
     for i in range( 0, 2 ):
         l = INf.readline()
     for i in range( 0, maxt ):
         l = INf.readline()
         l = l.strip()
         l = l.split()
         val = float( l[1] )
         if val < zbot01:
             val = 0.0
         elif val > zbot10:
             val = 1.0
         else:
             val = ( val - zbot01 ) / ( zbot10 - zbot01 )
         H.append( val )
     INf.close()
     data[ j, : ] = H

data_annual_min = np.zeros( [ len( INfdr ), 400 ] )
data_annual_max = np.zeros( [ len( INfdr ), 400 ] )
for j in range( 0, len( INfdr ) ):
    for i in range( 0, 400 ):
        data_annual_min[ j, i ] = np.min( data[ j, i*360 : i*360+360 ] )
        data_annual_max[ j, i ] = np.max( data[ j, i*360 : i*360+360 ] )

maxt, xincr, xtra = 400, 50, 20

mpl.rcParams[ 'font.sans-serif' ] = 'Calibri'
mpl.rcParams[ 'font.size' ] = 20

f = pyp.figure( figsize=[ 14, 10 ] )

s1 = f.add_subplot( 221 )
p = s1.boxplot( data_annual_min[ :, 100:200 ].T )
pyp.setp( p['boxes'], color='k' )
pyp.setp( p['whiskers'], color='k' )
pyp.setp( p['fliers'], color='k' )
#means = [ np.mean(x) for x in data_annual_min[ :, 100:200 ] ]
#pyp.scatter( np.arange( 1, 5 ), means, marker='o', edgecolor='k', facecolor='none' )
s1.set_ylabel( 'minimum fraction\nof wetland inundated' )
s1.set_ylim( [ -0.1, 1.1 ] )
s1.set_xticklabels( [] )
s1.set_title( '(a) Pre-watertable decline equilibrium\nperiod (i.e. 100-200 years
elapsed)', size=20 )

s2 = f.add_subplot( 222 )
p = s2.boxplot( data_annual_min[ :, 300:400 ].T )
pyp.setp( p['boxes'], color='k' )
pyp.setp( p['whiskers'], color='k' )
pyp.setp( p['fliers'], color='k' )
```

```
#means = [ np.mean(x) for x in data_annual_max[ :, 100:200 ] ]
#pyp.scatter( np.arange( 1, 5 ), means, marker='o', edgecolor='k', facecolor='none' )
s2.set_ylim( [ -0.1, 1.1 ] )
s2.set_xticklabels( [] )
s2.set_yticklabels( [] )
s2.set_title( '(b) Post-watertable decline equilibrium\nperiod (i.e. 300-400 years
elapsed)', size=20 )

s3 = f.add_subplot( 223 )
p = s3.boxplot( data_annual_max[ :, 100:200 ].T )
pyp.setp( p['boxes'], color='k' )
pyp.setp( p['whiskers'], color='k' )
pyp.setp( p['fliers'], color='k' )
#means = [ np.mean(x) for x in data_annual_min[ :, 300:400 ] ]
#pyp.scatter( np.arange( 1, 5 ), means, marker='o', edgecolor='k', facecolor='none' )
s3.set_ylim( [ -0.1, 1.1 ] )
s3.set_xticklabels( [ 0.0, 0.5, 1.0, 1.5, 0.0, 0.5, 1.0, 1.5, 0.0, 0.5, 1.0, 1.5 ] )
s3.set_xlabel( 'equivalent height of wetland\nsurface water addition (m)' )
s3.set_ylabel( 'maximum fraction\nof wetland inundated' )

s4 = f.add_subplot( 224 )
p = s4.boxplot( data_annual_max[ :, 300:400 ].T )
pyp.setp( p['boxes'], color='k' )
pyp.setp( p['whiskers'], color='k' )
pyp.setp( p['fliers'], color='k' )
#means = [ np.mean(x) for x in data_annual_max[ :, 300:400 ] ]
#pyp.scatter( np.arange( 1, 5 ), means, marker='o', edgecolor='k', facecolor='none' )
s4.set_ylim( [ -0.1, 1.1 ] )
s4.set_xticklabels( [ 0.0, 0.5, 1.0, 1.5, 0.0, 0.5, 1.0, 1.5, 0.0, 0.5, 1.0, 1.5 ] )
s4.set_yticklabels( [] )
s4.set_xlabel( 'equivalent height of wetland\nsurface water addition (m)' )

pyp.tight_layout()
pyp.savefig( 'Plot_boxwhisker_fraction_inundated.png', dpi=200 )
pyp.close( f )
```

C.3 Plot wetland salinisation risk metrics

```
import numpy as np
import matplotlib as mpl
import matplotlib.pyplot as pyp
from math import pi, sin
from os import path, getcwd

def file_len(fname):
    with open(fname) as f:
        for i, l in enumerate(f):
            pass
    return i + 1

def get_vals( INd, INf ):
    temp = []
    seep = []

    flen = file_len( path.join( getcwd() + INd, INf ) )
    I = open( path.join( getcwd() + INd, INf ), 'r' )
    check = '     CUMULATIVE VOLUMES      L**3       RATES FOR THIS TIME STEP
L**3/T\n'
    for i in range( 0, flen ):
        l = I.readline()
        if l == check:
            for j in range( 0, 8 ):
                l = I.readline()

            l = l.strip()
            l = l.split()
            temp.append( float( l[7] ) )

            if len( temp ) == 12:
                seep.append( np.array( temp ).sum() )
                temp = []
    I.close()
    seep = np.array( seep )

    emx, ecx, emy, ecy = -1.8, 1.3, 100., 140.
    ex = pi * emx + ecx
    ey = (( sin( ex ) * emy + ecy ) / ( 365.25 / 12. ) / 1000. * 0.7 ) * 100.0 * 10.0
    evap = np.ones( len( seep ) ) * ey
    metr = evap / ( evap + seep )
    return np.array( [ metr[0:100].mean(), metr[100:200].mean(), metr[200:300].mean(),
metr[300:400].mean() ] ).T

maxt, tincr, xtra = 144000, 18000, 1800

INd = [ '\\5p2a_SW0.0_FLT',
        '\\5p2b_SW0.5_FLT',
        '\\5p2c_SW1.0_FLT',
        '\\5p2d_SW1.5_FLT' ]

INf = 'output.dat'

data = np.zeros( [ 4, len( INd ) ] )
for s in range( 0, len( INd ) ):
    print INd[ s ]
    data[ :, s ] = get_vals ( INd[ s ], INf )

green   = [ 120.0 / 255.0, 190.0 / 255.0,   32.0 / 255.0 ]
lt_blue = [   0.0 / 255.0, 169.0 / 255.0 , 206.0 / 255.0 ]
dk_blue = [   0.0 / 255.0,  49.0 / 255.0 ,  60.0 / 255.0 ]
purple  = [ 159.0 / 255.0, 174.0 / 255.0 , 229.0 / 255.0 ]

xlabel = [ i[ -7: -4 ] for i in INd ]

mpl.rcParams[ 'font.sans-serif' ] = 'Calibri'
mpl.rcParams[ 'font.size' ] = 20
```

```
f = pyp.figure( figsize=[ 14, 6 ] )

s1 = f.add_subplot( 121 )
p = s1.bar( np.arange( -0.25, len( INd ) - 0.25, 1.0 ), data[ 1, : ], width=0.5,
color=lt_blue  )
s1.set_xticks( np.arange( 0, len( INd ) ) )
s1.set_xticklabels( xlabel )
s1.set_xlabel( 'equivalent height of wetland\nsurface water addition (m)' )
s1.set_xlim( [ -0.5, len( INd ) - 0.5 ] )
s1.set_ylim( [ -0.1, 1.1 ] )
s1.set_ylabel( 'salinisation risk metric' )
s1.set_title( '(a) Pre-watertable decline equilibrium\nperiod (i.e. 100-200 years
elapsed)', size=20 )

s2 = f.add_subplot( 122 )
p = s2.bar( np.arange( -0.25, len( INd ) - 0.25, 1.0 ), data[ 3, : ], width=0.5,
color=lt_blue )
s2.set_xticks( np.arange( 0, len( INd ) ) )
s2.set_xticklabels( xlabel )
s2.set_xlabel( 'equivalent height of wetland\nsurface water addition (m)' )
s2.set_xlim( [ -0.5, len( INd ) - 0.5 ] )
s2.set_ylim( [ -0.1, 1.1 ] )
#s2.set_ylabel( 'salinisation risk metric' )
s2.set_yticklabels( [] )
s2.set_title( '(b) Post-watertable decline equilibrium\nperiod (i.e. 300-400 years
elapsed)', size=20 )

pyp.tight_layout()
pyp.savefig( 'Plot_salinity_risk_metric.png', dpi=200 )
pyp.close( f )
```

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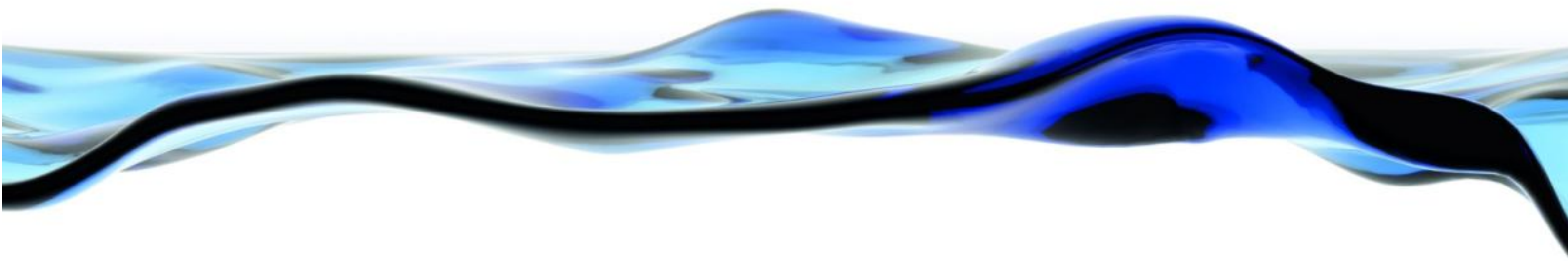

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