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Modelling the impacts of river stage manipulation on a complex river-floodplain system in a semi-arid region

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ABSTRACT

This paper investigates the complex interaction between a river and a saline floodplain in a semi-arid environment strongly influenced by groundwater lowering using a fully integrated physically-based numerical model. The main objective is to quantify the impacts of river stage manipulation on freshening of the shallow floodplain groundwater through bank storage. It is shown that river stage rises produce a relatively less saline floodplain aquifer with a larger freshwater lens. First, an increase in river stage reduces saline groundwater recharge to the floodplain. Second, the enhanced bank storage is able to freshen the groundwater near the river banks during high-flow pulses by mixing fresh water with saline groundwater. Overall, it was found that river stage manipulation may be considered as a short term salt management technique. However, if longer term strategies are required, it may be possible to implement these salt interception measures periodically.

Software

Name of software HydroGeoSphere
Developers R. Therrien (Universite Laval), R.G. McLaren, E.A. Sudicky, Y-J. Park (University of Waterloo)
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Hardware required Any Microsoft Windows based PC with sufficient RAM
Software required FORTRAN 95

1. Introduction

1.1. Floodplain salinization in arid and semi-arid regions

In arid and semi-arid environments, groundwater can be a major component of the water cycle (Ghazavi et al., 2012; Jolly et al., 2008). In these regions rainfall is typically seasonal, highly variable and significantly less than the evapotranspiration rate; therefore little, if any, diffuse groundwater recharge can occur. These factors create a natural tendency for salt accumulation in soils and groundwater. In floodplain environments, periodic natural overbank floods may prevent the development of soil and groundwater salinity (Meire et al., 2010; Restrepo et al., 1998; Zimmermann et al., 2006). Under natural conditions, arid and semi-arid floodplains occasionally experience periods of higher salinity as a consequence of high evaporation conditions and the variability of natural overbank floods, which provide dilution and flushing of the stored salt. However, due to the impacts of human population expansion and associated changes in land use, surface water regulation, and water resource depletion, arid and semi-arid floodplains, such as those in south-eastern Australia, are now often experiencing extended periods of low surface water flows and high soil salinity (Allison et al., 1990; Holland et al., 2013; Jolly et al., 2008). Consequently, the dynamic equilibrium of salinization and leaching is interrupted

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(Jolly et al., 1993). This can result in the reduced leaching of accumulated salt from root zones, thereby causing the dieback of environmentally important riparian vegetation, such as red gum (*Eucalyptus camaldulensis*) and black box (*Eucalyptus largiflorens*) and a decline in river water quality (Allison et al., 1990; Herczeg et al., 1993; Jolly et al., 1996; Peck and Hurle, 1973; Peck and Hatton, 2003).

1.2. Physical processes

Over the years there has been a growing awareness that the floodplain is a key area where a number of hydrogeological processes operate that have the potential to influence future outcomes of salinity management activities and river operations (Doble et al., 2006; Eslamian and Nekoueineghad, 2009; Evans et al., 2013; Ghazavi et al., 2012; Holland et al., 2009b). Therefore, studies have been conducted to develop models to quantify the impact of hydrological changes in river flow and floodplain ecology (Straatsma et al., 2013). A number of processes affect the flux exchange between a river and a floodplain in arid and semi-arid environments. These include rainfall, regional groundwater recharge, bank storage, evapotranspiration and groundwater extraction. Recharge from rainfall is often negligible in arid and semi-arid regions (Rassam et al., 2013).

Floodplains are generally topographically low in the landscape. Hence, the main recharge process in a floodplain aquifer is often groundwater flow from surrounding regional aquifers (Doble et al., 2006; Evans et al., 2013; Ghazavi et al., 2012). In Australia, the regional groundwater is usually naturally saline and often is the main source of solute movement towards the floodplain landscape. Groundwater recharge may be increased due to increased irrigation practices in the surrounding highland and this can lead to groundwater mounds (Fig. 1).

In arid and semi-arid regions, bank storage is an important process in the interaction between the surface and groundwater domains especially in rivers with high riverbed and riverbank hydraulic conductivities. Bank recharge represents a gain to the groundwater system. Three types of groundwater recharge were hypothesized by Jolly (2004) include bank recharge, diffuse recharge and localized recharge. Diffuse and localized recharge may occur during overbank flow which is not the focus of this paper. Bank storage is a dynamic phenomenon in which aquifer recharge occurs during periods of river stage rise followed by aquifer discharge to the river when the river stage reverts to a normal lower level (Ghazavi et al., 2012). During river stage recession, groundwater discharges to the river. The discharged groundwater usually has a solute concentration intermediate between that of the river and that of regional groundwater (McCallum et al., 2010). The net result of these processes at any point in space and time can lead to either a gaining or a losing river (Rassam, 2011). The observed response in rates of aquifer-floodplain exchange to changes in river stage strongly depends upon the state of connection between the two domains (Brunner et al., 2009). The significance of bank storage depends primarily upon the size of the river floodplain and its hydraulic and geometric properties (Doble et al., 2012; Knight and Rassam, 2007). Bank storage results in freshening of groundwater located near banks during high-flow pulses through the mixing of fresh river water with saline groundwater. For example, Holland et al. (2009a) showed that improvement in floodplain tree health was proportional to the extent of bank storage at different locations around a floodplain environment.

Groundwater evapotranspiration combines two processes: evaporation from groundwater lying close to the ground surface and transpiration from plants that use groundwater. In lowland gaining river-floodplain systems, groundwater flowing from the regional aquifer moves through the floodplain before either flowing to the river or being attenuated by evapotranspiration. A shallow groundwater presence within floodplains usually means that evapotranspiration rates are significant (Doble et al., 2006).

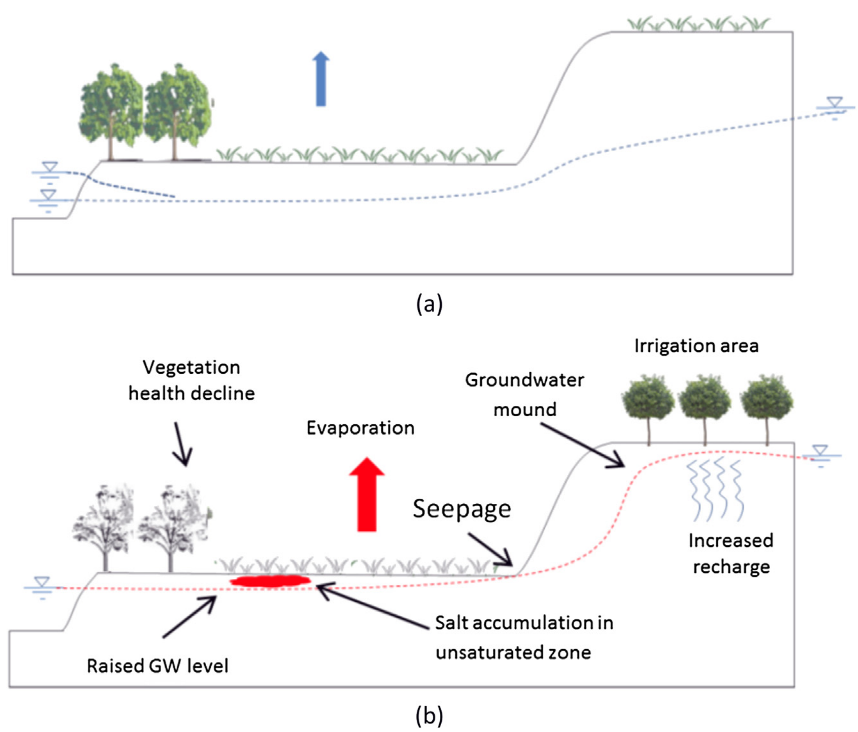

Fig. 1. Schematic of the SW–GW interaction across the Lower Murray River area before (a) and after (b) human-induced activities including weir and lock installations and irrigation practices.

Groundwater is often extracted via production wells for various purposes such as irrigation, water supply and groundwater lowering (as a salt interception measure). Depending on the size and hydrogeology of the floodplain and pumping rate of the production wells, groundwater lowering can significantly influence the SW–GW interactions (Rassam et al., 2013).

The unsaturated zone is often cited as the salt storage location during inter-flood periods. This is particularly the case in arid and semi-arid regions where the evapotranspiration rate is much higher than rainfall, which creates unsaturated solute storage zones in some areas of the floodplain. This is accelerated with changes in land use (such as irrigation recharge in the highland and floodplain), surface water regulation (such as raised groundwater level in the floodplain and reduction of high flow pulses) and water resource depletion (Fig. 1). Solute mass stored in the unsaturated zone appears to be correlated with the underlying groundwater salinity (Jolly, 2004).

1.3. Research challenges

Research and investigations into floodplain processes have mostly occurred over the last decade (Evans et al., 2013). Several studies have described some of the challenges of modelling SW–GW interactions in arid and semi-arid floodplains (Rassam et al., 2013). In addition, Rassam et al. (2013) highlighted the importance of incorporating SW–GW interactions into river management models that are used by water managers. In fact, uncertainty reduction will be very worthwhile since worldwide there is significant investment in water projects (Straatsma et al., 2013). The level of understanding of arid and semi-arid floodplain environments is still relatively basic, particularly in relation to SW–GW interactions in floodplains due to the highly complex nature of the floodplain environment (Eslamian and Nekoueineghad, 2009; Evans et al., 2013). McEwan et al. (2006) and Jolly et al. (2008) emphasized that while SW–GW interactions in temperate regions have been investigated in several studies (Hoehn and Scholtis, 2011; Kollet et al., 2010; Meire et al., 2010), floodplains in arid and semi-arid regions have received far less attention. However, many SW–GW modelling studies model water flow only and neglect solute transport (Crowe et al., 2004; Restrepo et al., 1998; Walton et al., 1996). Indeed, despite the fact that aquifer and surface waters are hydraulically interconnected, they are often modelled as two separate systems and are analysed independently (Schmid et al., 2006). The first published studies featuring fully coupled groundwater flow and solute modelling in relation to floodplain and river ecology have been undertaken relatively recently (Bauer et al., 2006a, 2006b; Langevin et al., 2005; Zimmermann et al., 2006). But, these studies generally did not model unsaturated zone processes, which are important to the prediction of ecological responses (Jolly, 2004). Another limitation of such studies is that many assumed steady-state GW and/or SW flow conditions, and these rarely exist in most arid and semi-arid floodplains, as these typically undergo periodic cycles of wetting and drying, resulting in transient SW–GW interactions (Jolly et al., 2008). Another issue is the high data requirements for this type of model as well as the lack of understanding of the key role of salinity in arid and semi-arid floodplains (Hart et al., 1991). There is a clear need to develop modelling capabilities for the movement of salt to, from, and within floodplains (Evans et al., 2013; Jolly et al., 2008). This can be addressed through developing a 3D physically-based fully integrated surface-subsurface numerical model with variable saturation and solute transport simulation capabilities (Alaghmand et al., 2013b).

1.4. Objective

Various management strategies can be used to maintain floodplain health. These include pumping saline groundwater, injection of fresh water, localised artificial flooding and environmental irrigation. These management measures require an understanding of SW–GW interaction at a fine scale, and the true ecological impact of land management decisions requires knowledge of the floodplain salinization risk. This paper investigates the complex interaction between a river (the Murray River in South Australia) and a saline floodplain (Clark's Floodplain) in a semi-arid area using a fully integrated physically-based numerical model featuring variable saturation and solute transport simulation capabilities. Clark's Floodplain is chosen as the study area due to the availability of sufficient recorded data which allows the development of a detailed unsteady-state (dynamic) model. However, the study period, which is from 1/01/2005 to 2/09/2010, is limited to just under five years and the scenarios are representative of transient behaviour that may not illustrate the system moving to new equilibrium positions. In addition, the study period corresponds to very dry conditions and so rainfall fluxes and river stage variations could have been much less than long-term variations. The main objective is therefore to quantify the impacts of river stage manipulation on freshening of the shallow floodplain groundwater through bank storage. The river-floodplain system is complex as the floodplain aquifer is strongly influenced by a Salt Interception Scheme (SIS) that involves groundwater lowering. Hence, various scenarios are defined to understand the combined and individual impacts of river stage manipulation and groundwater lowering on the flow and the solute dynamic of the floodplain aquifer. The hypothesis that is tested here is that higher river stages lead to a relatively less saline floodplain aquifer by increasing the fresh river water flux to the floodplain aquifer and reducing the saline groundwater flux from the highland to the floodplain aquifer.

2. Materials and methods

A fully integrated surface-subsurface numerical model is developed for Clark's Floodplain, as described in detail below. The model is calibrated to data from a time period that includes operation of the SIS production wells. Scenarios are used to determine the relative impacts of river stage manipulation on water and solute balances within the floodplain aquifer.

2.1. Study site

Clark's Floodplain is located on the Lower Murray River in South Australia (34°21′S, 140°34′E) (Fig. 2) next to the Bookpurnong Irrigation District. The study site is located in a semi-arid region of South Australia, with annual rainfall varying between 200 and 300 mm and annual areal potential evaporation of 1800 mm. Data from Loxton meteorological station shows that local evaporation rates were continuously higher than rainfall depths between 2005 and 2010 (BOM, 2013). At the study site the Coonambidgal Clay (typically consisting of clays and silts) ranges from 2 to 7 m thick, while the Monoman Formation (coarse-grained quartz sands) is approximately 7 m thick (Fig. 2). The highland adjacent to the floodplain consists of a layer of Loxton Sands (Upper and Lower units) up to 35 m in depth. The whole area is underlain by the Loxton Sand and Bookpurnong Beds, the latter acts as an aquitard basement to the shallow aquifer that includes the Monoman Formation and Loxton Sands (AWE, 1999; Barnett et al., 2002; Doble et al., 2006). For further details on the hydrogeology of the study site the reader is referred to Doble et al. (2006) and Alaghmand et al. (2013a).

The increased groundwater recharge contributed by the Bookpurnong Irrigation District has locally raised the water-table in the Loxton Sands (Telfer and Overton, 1999). The increased groundwater gradient between the Loxton Sands aquifer and the Murray River has led to greater salt flux from the saline regional aquifer into the floodplain and river. Groundwater salinity in the floodplain has also increased due to a lack of floods that could potentially freshen the groundwater via bank storage. Black box and red gum tree communities have been most affected by the salinization of the floodplain (Doble et al., 2006). Groundwater salinity in the Loxton Sands–Monoman Formation aquifer is typically in excess of 50,000 $\mu S\ cm^{-1}$, while irrigation recharge salinity is typically 8000 $\mu S\ cm^{-1}$. In an effort to mitigate such impacts, salt interception schemes (SISs) have been implemented at various sites along the Lower Murray River, which intercept saline groundwater before it reaches the river (White et al., 2009). Two of the SIS production wells, 32FP and 34FP, are located at the study site and these have significant impacts on the SW–GW interactions.

Meteorological, hydraulic and hydrogeological data for the Clark's Floodplain site are well-documented for the modelled time period (1/1/2005–2/09/2010) and

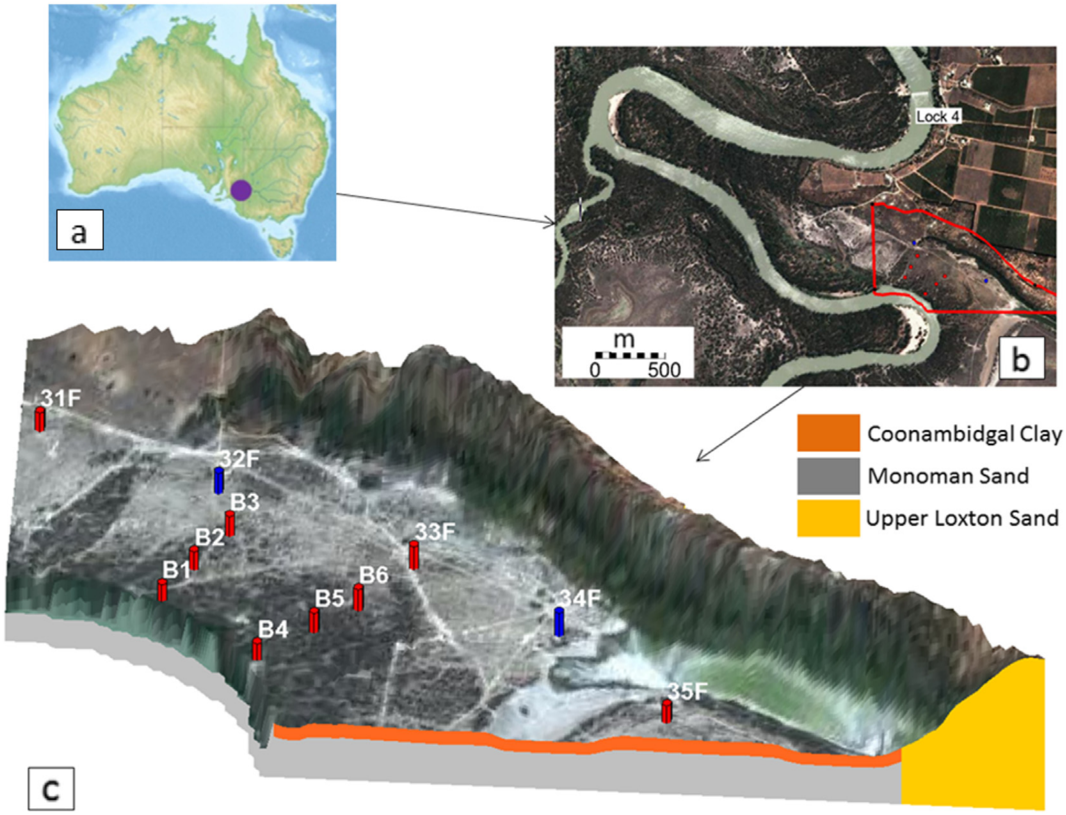

Fig. 2. a: Location of Clark's floodplain in Australia (shown in purple), b: Perimeter of the geometry model (shown in red), c: 3D visualization of the geometry of the study site including the soil types and observation (B1, B2, B3, B4, B5, B6, 31F, 33F and 35F) and SIS production wells (32F and 34F) (Z magnification = 8). The cover image is adopted from GoogleMaps. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

were used to inform initial and boundary conditions for the model. The time period was chosen to include the non-flooding conditions and a flood event which occurred in 2010. Recorded meteorological data, such as rainfall and potential evaporation, were obtained from Loxton meteorological station (BOM, 2013). Murray River flows at the study site were obtained from Lock 4 water level station, which is situated immediately upstream of Clark's Floodplain (WaterConnect, 2013) (Fig. 2). River salinity observations were obtained from Clarke's Sandbar and Rilli Island stations located upstream and downstream of the study site, respectively (WaterConnect, 2013). Floodplain potentiometric head and solute concentration observations were obtained from groundwater observation wells located along two transects that extend from the river to the floodplain perimeter: B1, B2 and B3 on transect 1; and B4, B5 and B6 on transect 2 (Fig. 2) which were adopted from Berens et al. (2009) and Holland. Built in 2004 and 2005, the observation wells were designed to monitor groundwater levels and salinity at depths from 2 m up to 10 m, since salinity is anticipated to occur at the top of water table (Berens et al., 2009).

2.2. Numerical model

Characterisation of near-river-aquifer systems is complex because of the nature of SW–GW interaction processes, and the uncertainty of land cover and aquifer properties, which can produce significant errors in hydrodynamic models outputs (Eslamian and Nekoueineghad, 2009; Sophocleous, 2010; Straatsma et al., 2013). This can be addressed using a fully-integrated, physically-based numerical model (Alaghmand et al., 2013b). Due to the required capabilities, the available observed input data, the scale of the study, and the required robustness and stability of the numerical methods, the HGS model (Therrien et al., 2006) was selected for this research. HGS is a three-dimensional numerical model describing fully-integrated surface and subsurface flow and solute transport. HGS models the flow of water through unsaturated porous media by numerical solution of the Richards equation. The van Genuchten (1980) or Brooks-Corey (Brooks and Corey, 1964) relationships are used to relate pressure head to saturation and relative hydraulic conductivity. Surface water flow is modelled using two-dimensional depth-averaged flow. Saturated groundwater flow is modelled by numerical solution of the groundwater flow equation. Two surface and subsurface coupling approaches are available in HGS, namely the common node approach (based on continuity of hydraulic head between two domains) and the dual node approach (based on a first-order exchange coefficient), with the latter being used in this study. Transpiration from vegetation occurs within the root zone of the subsurface and is a function of the leaf area index (LAI), nodal water (moisture) content and a root distribution function (RDF) over a prescribed extinction depth. Evaporation from the soil surface and subsurface soil layers is a function of nodal water content and an evaporation distribution function (EDF) over a prescribed extinction depth. The model assumes that evaporation occurs along with transpiration, resulting from energy that penetrates the vegetation cover. For further details on the code and a recent software review the reader is referred to Therrien et al. (2006) and Brunner and Simmons (2012). HGS requires pre- and post-processor tools in order to handle input preparation (complex topography and grid) and visualization of the outputs. In this study, Grid Builder (McLaren, 2005) and Groundwater Modelling System (GMS) (AquaVeo, 2011) were used to generate the model grid. GMS was also used to visualize and interpret the model outputs. In this study, HGS used the control volume finite element approach to solve surface and subsurface flow and transport. The model was a transient model setup for a period of 2070 days (from 1/01/2005 to 2/09/2010) using an initial time step of 0.1 days, a maximum time step of 1 day and a maximum time step multiplier of 1.25. The model solves non-linear equations for variably-saturated subsurface flow, surface flow and solute transport. To solve the non-linear equations, HGS uses the Newton–Raphson linearization method. Newton iteration parameters include Newton maximum iterations (25), Jacobian epsilon (10.0 d^{-5}), Newton absolute convergence criteria (1.0 d^{-5}), Newton residual convergence criteria (1.0 d^{-3}) and flow solver maximum iterations (1.0 d^{5}).

2.3. Model set up

2.3.1. Geometry grid

The model domain perimeter is shown in Fig. 3. The model spatial discretisation is based on a LiDAR Digital Elevation Model of the study site with a 10 m grid resolution. The resulting grid consisted of 78,624 nodes and 143,500 elements. As shown in Fig. 3, the geometric grid covers 61.3 ha of Clark's Floodplain from the floodplain slope break to the Lower Murray River main channel. This includes two SIS production wells (32F and 34F) and nine observation wells (Fig. 3). In this case, the length of the river bank is 570 m and the distance from the river bank to the SIS wells varies between 480 m and 650 m.

2.3.2. Parameters

Three soil types were represented, namely a continuous 10 m-thick layer of Monoman Formation sand, overlaid by a spatially variable, 2 to 6 m-thick layer of semi-confining heavy Coonambidgal Clay and Upper Loxton Sand in the adjacent

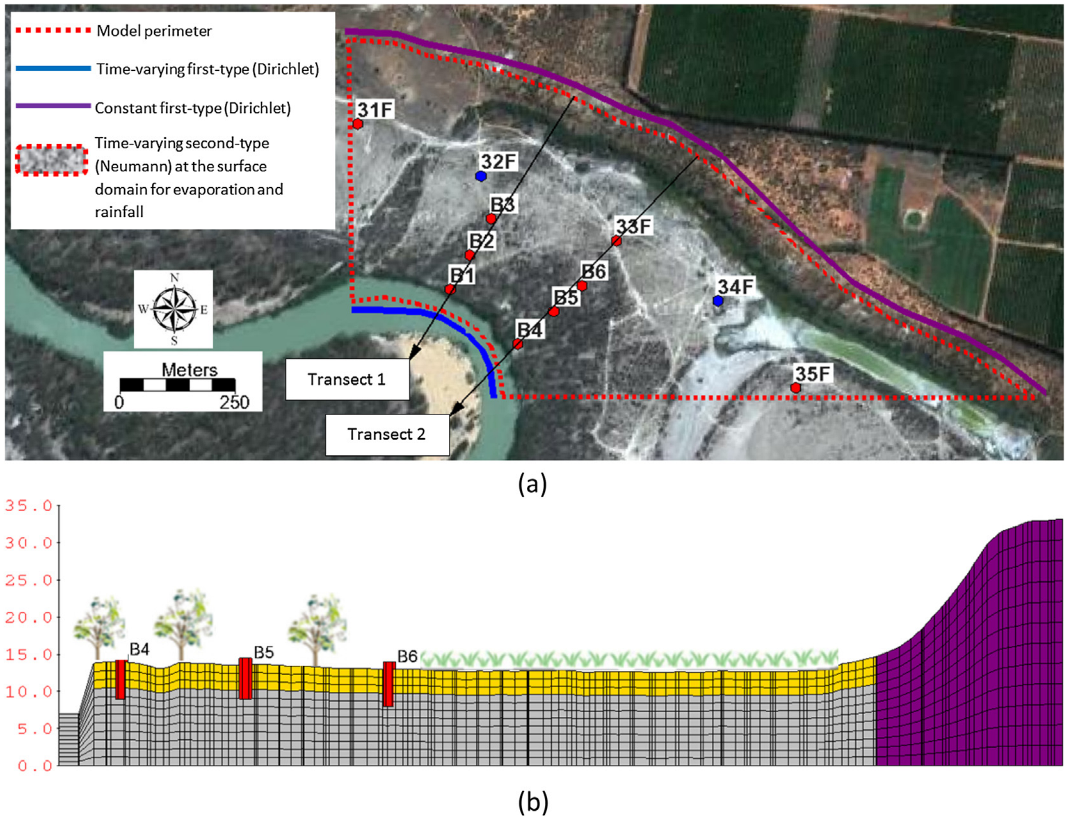

Fig. 3. a: Configuration of the model boundary conditions, b: Configuration of the vegetation and soil layers of Clark's Floodplain along transect 1 (Z magnification = 3). Observation wells are shown as black columns.

highland (Fig. 3). Soil properties, such as hydraulic conductivity (isotropic), porosity, residual saturation and specific storage, were obtained from Carsel and Parrish (1988) and Doble et al. (2006). Van Genuchten function parameters (n and alpha) (van Genuchten, 1980) were adopted from Jolly et al. (2008) who adjusted and proposed these parameters for the Lower Murray River soil types. Longitudinal and transverse solute dispersivity values were estimated through model calibration. The hydraulic properties of the surface domain (river bed and floodplain corridor) have significant differences and so these were divided in the model into main channel (river) and floodplain. Furthermore, the vegetation coverage of the floodplain was divided into two different categories (*Eucalyptus* trees and grass) and evapotranspiration parameter values for both categories were adopted from Hingston et al. (1997), Banks et al. (2011) and Verstrepen (2011). Table 1 summarises the parameters values used in the numerical model.

2.3.3. Boundary conditions

The locations where boundary conditions were specified are given in Fig. 3a. Two types of boundary conditions were used in the model including first-type (Dirichlet) boundaries of prescribed head/concentration and second-type (Neumann) boundaries of prescribed flow/solute flux. In the subsurface (porous media) domain, a constant first type (Dirichlet) boundary condition of 12 m AHD (Australian Height Datum) constant head was specified at the north-eastern part of the domain. This condition was adopted based on potentiometric contours (AWE, 2013). The observed river levels for the surface domain were set at the river side of the model using a time-varying first-type (Dirichlet) boundary condition. In this regard, the observed water levels downstream of Lock 4 were applied to the river nodes of the model (WaterConnect, 2013).

To represent the solute boundary conditions, a first-type (Dirichlet) constant concentration boundary condition was assigned. The observed groundwater concentrations at the observation wells in the river and the floodplain ranged from 300 μS cm^{-1} to 50,000 μS cm^{-1} (Holland et al., 2013). Hence, constant values were applied at the subsurface outer boundary (representing regional groundwater in the highland aquifer) and the river nodes accordingly. Two SIS production wells (30F and 32F) were represented in the model using recorded pumping rates and durations obtained from Berens et al. (2009).

Rainfall was modelled for the entire model surface domain beginning on day 1 using a time-varying second-type (Neumann) boundary condition according to recorded data (BOM, 2013). Evapotranspiration was dynamically modelled as a combination of evaporation and transpiration processes by removing water from all model cells of the surface and subsurface flow domains within the defined zone of the evaporation and root extinction depths.

2.3.4. Initial conditions

The initial conditions for the calibration model were obtained from a steady-state flow and transport model that represented the status of the river-floodplain system prior to the study period (2005). The initial model was ran for long enough (30 years) to reach the equilibrium condition (Barnett et al., 2012). Hydraulic head and solute concentration outputs from the initial model compared favourably with observations from six observation wells on Clark's Floodplain recorded in 2005 which was available through Holland et al. (2013). Also, the status of the solute concentration distribution at the beginning of the study period was checked with the general observed solute distribution pattern in the floodplain. This can be considered as two zones: a relatively fresh GW zone within 50 m distance of the river banks (B1: 6500 μS cm^{-1} and B4: 1200 μS cm^{-1}); and a saline zone (B2: 53,000 μS cm^{-1}, B3: 54,000 μS cm^{-1}, B5: 50,900 μS cm^{-1} and B6: 52,000 μS cm^{-1}) for the rest of the floodplain (Fig. 6c).

2.4. Coupled flow and transport calibration

Calibration was undertaken using an iterative trial-and-error method. In order to minimise the uncertainty associated with parameters such as hydraulic conductivity, porosity, dispersivity (longitudinal and transverse) and leaf area index, these were altered within known ranges and reasonable limits in order to achieve an acceptable match to observations of hydraulic head and solute concentrations pattern. Two different approaches were employed for the flow and solute calibrations. While, the aim of the calibration process for flow is to match the absolute groundwater heads at the observation wells, the solute is calibrated to the observed concentration patterns. This is because concentration patterns are much more sensitive to local-scale geological heterogeneity than are hydraulic heads, and models may have difficulty reproducing the concentrations or their temporal variability at single observation wells (Barnett et al., 2012).

2.5. Numerical model performance evaluation

In order to obtain a reasonable evaluation of the numerical model performance, several factors need to be taken into account. These may include the field of application, characteristics of the model, available observed data, information and knowledge of the problem, and the specific objectives of the modelling exercise

Table 1
Parameter values of the model for the study site.

Model parameter	Value			Units
Subsurface domain	Monoman Sand	Coonambidgal Clay	Upper Loxton Sand	
Porosity	35	60	40	%
Hydraulic conductivity	20	0.1	10	m d^{-1}
Specific storage	1.6×10^{-4}	2.0×10^{-3}	1.6×10^{-4}	m^{-1}
Evaporation limiting saturation (min)	0.05	0.25	0.15	
Evaporation limiting saturation (max)	0.9	0.9	0.9	
Longitudinal dispersivity	3	3	3	m
Transverse dispersivity	0.3	0.3	0.3	m
Residual water content	0.04	0.04	0.04	
Alpha	1.69	0.28	0.8	m^{-1}
n	8.25	2.52	3.6	
Evapotranspiration		Eucalyptus	Grass	
Tree canopy evaporation		4.5×10^{-4}	4.0×10^{-4}	m
Evaporation extinction depth defined by quadratic decay Evaporation distribution function		1	1	m
Transpiration extinction depth defined by quadratic decay Root distribution function		5	0.5	m
Leaf area index		0.5	0.5	m^2 m^{-2}
Transpiration fitting parameter c1		0.3	0.6	
Transpiration fitting parameter c2		0.2	0	
Transpiration fitting parameter c3		1	1	
Transpiration limiting saturation (at wilting point)		0.29	0.29	
Transpiration limiting saturation (at field capacity)		0.56	0.56	
Transpiration limiting saturation (at oxic limit)		0.85	0.75	
Transpiration limiting saturation (at anoxic limit)		0.95	0.9	
Initial interception storage		3.0×10^{-4}	4.0×10^{-4}	m
Surface domain		River	Floodplain	
Friction (*x*-plane)		5.0×10^{-3}	5.0×10^{-2}	
Friction (*y*-plane)		5.0×10^{-3}	5.0×10^{-2}	
Rill storage height		1.0×10^{-3}	1.0×10^{-2}	m
Coupling length		1.0×10^{-2}	1.0×10^{-2}	m
Obstruction storage height		0	1.0×10^{-3}	m

(Bennett et al., 2013; Jakeman et al., 2006; Matthews et al., 2011). Moreover, environmental models typically have multiple interacting drivers with uncertain properties (Bennett et al., 2013; Rassam et al., 2013). Hence, multiple evaluation metrics need to be used for a comprehensive evaluation of the numerical model. Otherwise, a single performance criterion approach may lead to counterproductive results such as favouring models that do not reproduce important features of a system (Bennett et al., 2013; Gupta et al., 2012).

A number of quantitative approaches are available to assess the model performance (Bennett et al., 2013). For instance, direct value comparison methods aim to test whether the modelled values show similar characteristics as a whole to the observed values. In this case, the means of the modelled and observed data sets are compared and expressed as Means Differences in Table 2. Clearly the ideal value would be zero. Furthermore, some model performance evaluation methods, such as the residual method, involve coupling observed and modelled values. In the residual method the difference between modelled and observed data are calculated. Of the many possible numerical calculations on model residuals, Mean Square Error (*MSE*) and Root Mean Squared Error (*RMSE*) are considered here. The ideal value for both of these metrics is zero. Another model performance evaluation metric involves preserving the data patterns. This method tests the ability of the model to preserve the patterns of observed and modelled data. The Coefficient of Determination (r^2) is one of the metrics in this category which indicates how variation of one variable is explained by a second variable. This is commonly used to measure the efficiency of a model and values range between 0 and 1. Another Coefficient of Determination which is popular in hydrologic modelling is the Nash–Sutcliffe Model Efficiency (*NSE*) (Nash and Sutcliffe, 1970). This ranges from $-\infty$ to 1 and indicates how well a model explains the variance in the observations, compared with their mean as the prediction. The ideal value for both of these metrics is one. A detailed description of the qualitative and quantitative methods of characterising performance of environmental models is provided by Bennett et al. (2013).

In addition to the above mentioned quantitative model performance evaluation methods, visual performance measures have been developed to mimic how the eye evaluates proximity between observed and modelled values (Ehret and Zehe, 2011; Ewen, 2011). This type of qualitative method avoids traducing model errors simply in terms of difference of magnitude and, also includes time shifts. In fact, qualitative assessments are important in complex models as they enable the modeller to sketch out trends and system behaviour rather than producing actual values for variables (Bennett et al., 2013).

2.6. Scenarios

The calibrated model represents the observed SW–GW interaction from 1/1/2005 to 2/09/2010. During the study period the SIS was in operation and there was no river manipulation (hereafter referred to as the Only-SIS scenario). To predict SW–GW interactions induced by river stage manipulation, the calibrated model was re-run while imposing various river stage elevations. Hence, twelve hypothetical river stage manipulation scenarios were defined for water stage increases of 0.5 m, 1.0 m and 1.5 m and for a decrease of 0.5 m. Each of these water stage changes was modelled for 1 month, 2 months, and 3 months in each year. Model simulations covered the period between 1/1/2005 and 2/9/2010 (2070 time steps). The response of the floodplain aquifer to the various scenarios was observed in terms of hydraulic heads and solute dynamics. River stage manipulation was not the only stress on the model during the study period as groundwater lowering also occurred via the operation of the SIS production wells. Hence, one scenario without groundwater lowering (hereafter referred to as the No-salt management scenario) was included as well. Fig. 4 shows the hydrographs for the manipulated river stage elevations for the defined scenarios.

3. Results and discussion

Results of the numerical model are discussed in five sections. First the results of the calibrated model are demonstrated and discussed along with No-salt management scenario. This is followed by discussion of the model results in terms of water balance and solute balance for the defined scenarios. Then, solute mass in the unsaturated zone is analysed and finally, the ecological implications of river stage manipulation are discussed.

3.1. Calibrated model

The calibrated model represents the observed behaviour of the river-floodplain system in terms of water and solute dynamics over the period 1/01/2005 to 2/09/2010. The numerical model performance in terms of groundwater head was tested both qualitatively and quantitatively. The observed and modelled series of hydraulic heads and solute concentrations at observation wells B1, B2, B3, B4, B5 and B6 were also compared visually (Fig. 5). Moreover, quantitative evaluation was undertaken using the model performance evaluation metrics discussed in Section 2.4 including Means Difference, Coefficient of Determination (r^2), Mean Sum of Error (MSE), Root Mean Squared Error (RMSE) and Nash–Sutcliffe Model Efficiency (NSE) (Table 2). Considering Fig. 5 and Table 2, it appears that the numerical model properly reproduced the observed groundwater dynamic during the study period.

The modelled groundwater salinity distribution result displays the presence of freshwater along the eastern margin abutting the river channel and a saline zone in the rest of the floodplain aquifer (Fig. 6a). The observed groundwater salinities at the location of the observation wells are shown in Fig. 6c (Holland et al., 2013). This shows that the observation wells in the vicinity of the river bank (B1 and B4) have significantly lower salinity than the other four observation wells located further away on the floodplain (B2, B3, B5 and B6). Moreover, an EM31 survey was conducted in November 2007

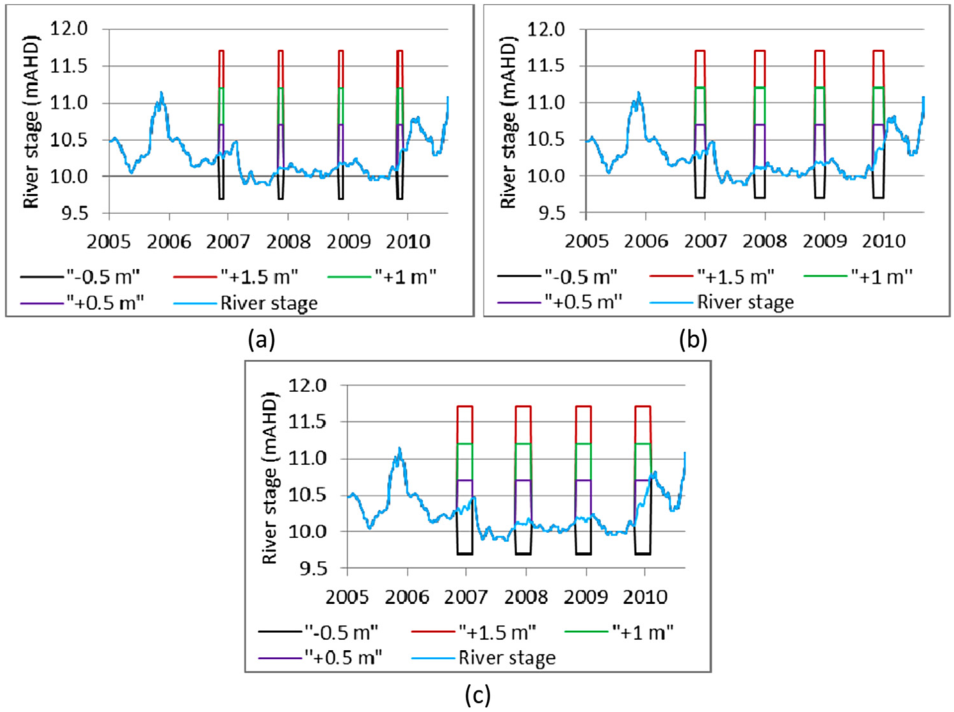

Fig. 4. Time-varying river stage boundary conditions for scenarios featuring river stage rise durations of (a) one month, (b) two months, and (c) three months.

and reported by Berens et al. (2009). Depending on subsurface conductivity, the EM31 has a limited penetration depth of approximately 4–6 m, and yields a bulk conductivity representation of that shallow interval. Variables that may typically influence the results of the EM31 survey include groundwater depth and salinity, variations in soil moisture and salinity, and the clay content. However, with the Murray River floodplains consisting mainly of sands and localised clays of similar porosity, the water content in the saturated environment is most likely consistent, leaving salinity as the main driver of conductivity. Hence, the EM31 results can be a proper indicator of groundwater salinity at the study site. This is shown in Fig. 6b. The modelled groundwater salinity distribution (Fig. 6a) and conductivity distribution obtained from the EM31 survey in November 2007 (Fig. 6b) present a good agreement. Overall, it is confirmed that the calibrated model is able to reproduce the solute dynamic of the surface-groundwater interaction processes in an acceptable manner as it is consistent with the observed data.

Fig. 7 shows the groundwater balance for Only-SIS and No-salt management scenarios. As rainfall recharge is unlikely to happen at the study site, the river and regional groundwater are the dominant recharge features in both models. As a general trend, the bank storage strongly responds to the river stage fluctuation, while regional groundwater recharge is a function of the operation of the SIS production wells. Both bank storage and regional groundwater are larger during the operation of the SIS production wells (Fig. 7a). This shows the boosted hydraulic gradient towards the SIS production wells during their operation. In Only-SIS model, when the SIS is shut down for a short period (November 2006–April 2007), the dominant recharge feature is regional groundwater. In this period the bank storage is at a minimum due to the reversed hydraulic head towards the river. On the other hand, in No-salt management scenario only two major river bank recharges occur and these are responses to the two high flows in December 2006 and January 2010.

The main discharge processes are evapotranspiration, bank discharge and groundwater extraction via the SIS in Only-SIS scenario. In No-salt management scenario, water discharge via evapotranspiration is slightly higher than Only-SIS scenario. This is because of the relatively shallower groundwater table in No-salt management model. Generally, during the operation of the SIS production wells, the floodplain aquifer has a gaining regime while in No-salt management scenario it is mostly losing. The discharge via the river bank occurs continuously in No-salt management scenario. This is due to the higher groundwater table in the floodplain aquifer except during the two river high flows. But in Only-SIS scenario, only two major groundwater discharges were observed. The first of these was prior to the commencement of the SIS operation when the river stage was lower than the groundwater table. The second was three months after the SIS was shut down. Both of the major groundwater discharges were diminished when the SIS commenced (July 2005 and May 2007). These are consistent with the observed and modelled groundwater head dynamics, as shown in Fig. 5. It seems that the SIS operation lowers the groundwater table and enhances fresh river water recharge to the floodplain aquifer on one side and saline regional groundwater recharge on the other side of the production wells.

Fig. 8 illustrates the solute mass balance for Only-SIS and No-salt management scenario. The results show that the stored solute mass in the floodplain aquifer is reduced during the study period of Only-SIS scenario by 4% (Fig. 8a). Although without the SIS operation, the stored solute mass would have increased by 5% (Fig. 8b). It seems that groundwater lowering via the SIS production wells may lead to a less saline floodplain aquifer. This happens through two mechanisms, namely the extraction of some portion of the saline groundwater (Fig. 9) and the reversal of the hydraulic head towards the floodplain aquifer. The 5–15 ton d^{-1} solute mass lowering from the wells is not substantial in comparison with the total stored solute mass in the floodplain aquifer (around 60,000 ton in Fig. 8). Therefore, the main mechanism is reversing the hydraulic head towards the floodplain aquifer. In fact, the SIS production wells create a divide which stops saline water from reaching the floodplain by lowering the groundwater table. In other words,

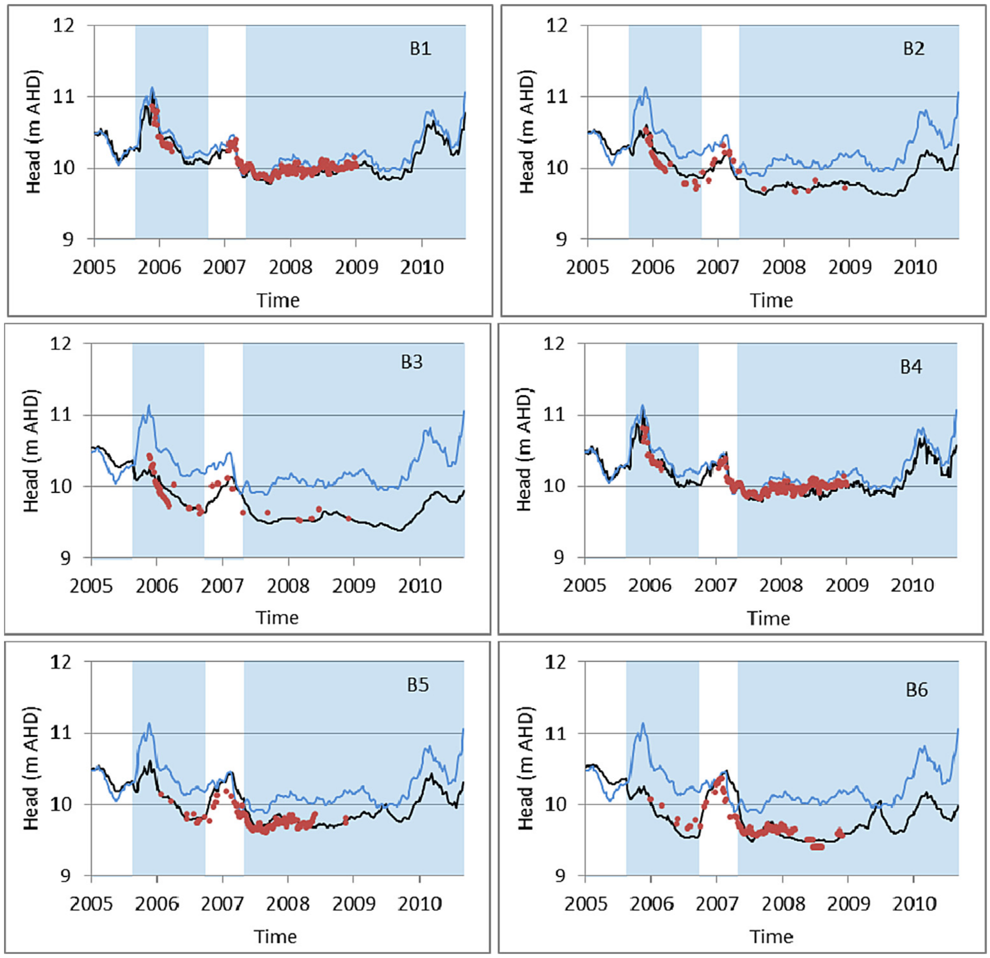

Fig. 5. Modelled and observed groundwater heads at the observation wells. River stage and modelled and observed groundwater heads are shown as blue lines, black lines and red dots, respectively. The light blue pattern represents the periods during which the SIS production wells were in operation. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

groundwater lowering enhances the fresh water lens on one side and keeps the saline groundwater on the other side. But, in the absence of the SIS production wells, there is no process to prevent the saline groundwater from reaching the river and the only discharge process is via the river bank. Hence, the recharged solute mass from regional groundwater is stored in the floodplain aquifer or discharges to the river and leads to a more saline floodplain. Fig. 10 shows the impact of the SIS production wells operating along transect 1.

Regarding Fig. 8a, two major solute mass discharges to the river bank are observed in June 2005 and March 2007 (up to 2 ton d^{-1}). Considering that approximately 800 ton d^{-1} of river salt load was recorded at Lock 4 just upstream of the study site during the same periods (WaterConnect, 2013), the solute mass discharge via the river bank should not have a significant impact on the river water quality. This indicates that the dominant solute mass discharge process is saline groundwater extraction via the SIS production wells. In addition, in No-salt management scenario (Fig. 8b), saline groundwater constantly discharges to the river at a higher rate (up to 4 ton d^{-1}). This would present a risk to river water quality over the long term.

3.2. Water balance

The river stage manipulation scenarios are now considered. One of the main starting points for analysis of the flow dynamics in the surface-groundwater system is the water balance. Hence, the outputs from the numerical model for each round of river stage manipulation (hereafter referred to as *trials*) that are considered here include groundwater table dynamics, change in water storage in the floodplain aquifer (state of gaining or losing floodplain), flux exchange between the two domains (bank storage) and recharge from regional groundwater.

The dynamics of the GW heads at the observation wells along transect 1 (wells B1, B2 and B3) are shown in Fig. 11. As expected, a rising river stage creates higher gradients from the river to the floodplain aquifer. Obviously, the GW dynamic is much more enhanced near the river bank rather than further away. This is noticeable in Fig. 11a and b where observation well B4 shows a greater response to the river stage manipulations. Also, longer operation of the SIS wells has a greater effect on GW heads. This can be seen in Fig. 11c whereby the same river stage rise scenario (here 1.5 m) with a longer trial duration shows a higher average groundwater hydraulic head in the floodplain aquifer at observation well B6. In other words, the extent of the floodplain aquifer response increases as the duration of SIS operations increases.

Fig. 12a shows the change in water storage in the floodplain aquifer during the study period. When this parameter is positive it represents a gaining floodplain while a negative value indicates that the floodplain has a losing regime. The change in water storage generally depends on the conductance (which does not vary over time) and the time-varying head gradient between the river and

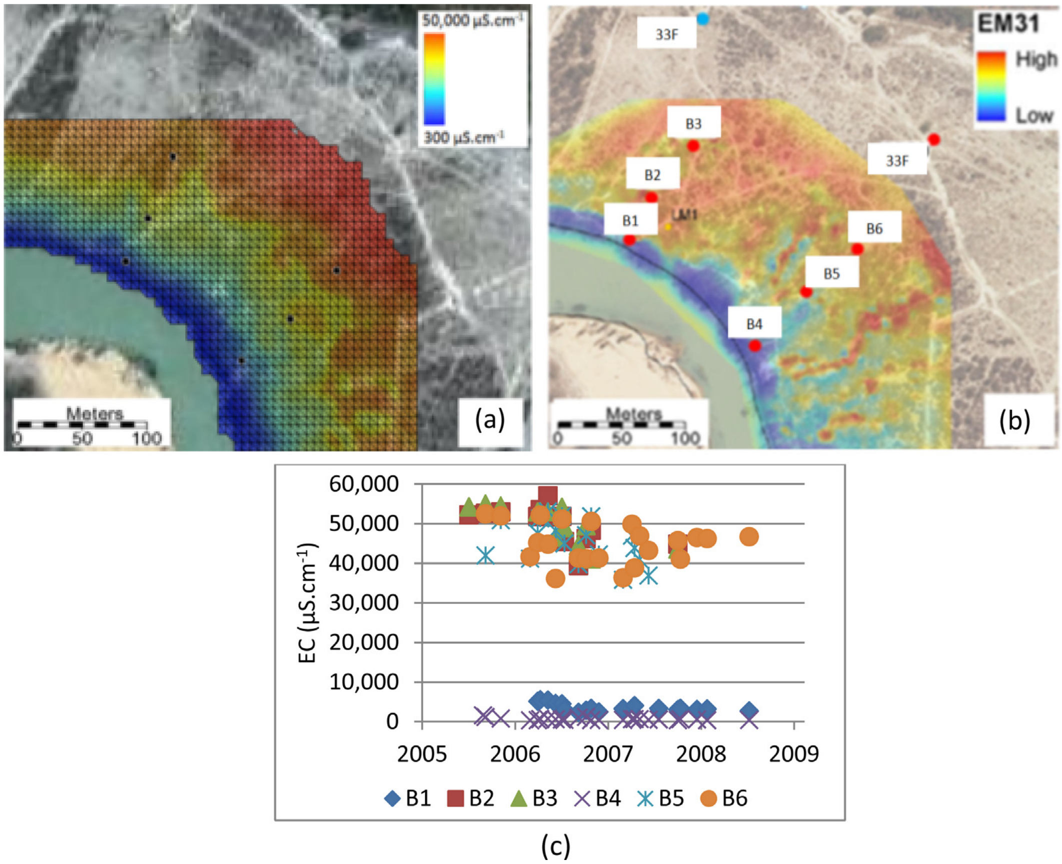

Fig. 6. a: Modelled groundwater salinity distribution (November 2007, time step 650 days), b: Conductivity distribution, EM31 survey in November 2007 (Berens et al., 2009), c: Recorded groundwater salinity during the study period (Holland et al., 2013).

groundwater. According to Fig. 9, the floodplain aquifer was approaching a gaining condition due to the river stage rise just before commencement of the SIS operation in July 2005. But, commencement of groundwater extraction via the SIS bores quickly formed a losing floodplain. This explains the formation of the pumping drawdown cone. The floodplain aquifer was in a losing condition until the first trial in November 2006, except for a short period in December 2005 corresponding to a high river flow.

The response of the floodplain aquifer to each of the trials is evident in Fig. 12a. As a general pattern, each rising scenario leads to a gaining floodplain during each trial and a losing floodplain after the trials. Clearly, the magnitude of the response to each trial is proportional to the height of the river stage rise. In contrast, the 0.5 m drop scenario creates a losing floodplain during the trial and a gaining one after the trial.

Due to the complexity of the study site, the floodplain response to each of the four trials is different. The first trial is coincident with the period during which the SIS was shut down. Hence, one of the main discharge components was absent. This created the most enhanced gaining floodplain condition (maximum value in Fig. 12a). In fact, in this period the floodplain aquifer was recharged from the river and from regional groundwater. However, during the 2nd trial a sudden decrease in the pumping rates of the SIS production wells occurred (from 5.5 l s^{-1} to 2.2 l s^{-1}). This led to the lowest gaining condition among the four trials. The responses to the 3rd and 4th trials are the result of both the operation of the SIS production wells and river stage manipulation.

For the 0.5 m drop scenario, the highest losing and gaining conditions happened during the 2nd and 4th trials, respectively. During the 2nd trial the hydraulic head increased towards the river due to a decrease in the pumping rate of the SIS production wells. For the 4th trial, the gaining condition was coincident with a river high flow. Moreover, the highest losing condition (minimum values in Fig. 12a) occurred just before resumption of the SIS operation, but this was not due to river stage manipulation. This is partly attributed to the resumption of the SIS production wells and partly to the river stage decrease during that period.

Fig. 12b shows the flux exchange between the river and floodplain aquifer during the study period for the 3 month scenarios. This shows that the increase of bank storage is proportional to the rise in river stage. For example, as more water enters the floodplain aquifer, consequently more water returns back to the river during the river stage recession and this results in a greater flux for the 1.5 m scenario than for the other scenarios. This is due to the head difference that is formed with each river stage manipulation. After each trial, a minimum bank storage can be observed. It can also be seen that the volume of water that enters the aquifer from the river during the trial is greater than that which is subsequently

Table 2
Model performance evaluation metrics (*Means Difference*; *MSE* = Mean Square Error; *RMSE* = Root Mean Squared Error; r^2 = Coefficient of Determination; *NSE* = Nash–Sutcliffe Model Efficiency coefficient).

Observation Well	*Means Difference* (*m*)	*MSE* (*m*)	*RMSE* (*m*)	r^2	*NSE*
B1	0.05	0.054	0.067	0.91	0.76
B2	0.07	0.075	0.088	0.87	0.71
B3	0.12	0.080	0.091	0.85	0.66
B4	0.06	0.044	0.058	0.89	0.77
B5	0.10	0.031	0.041	0.83	0.63
B6	0.05	0.048	0.061	0.81	0.61

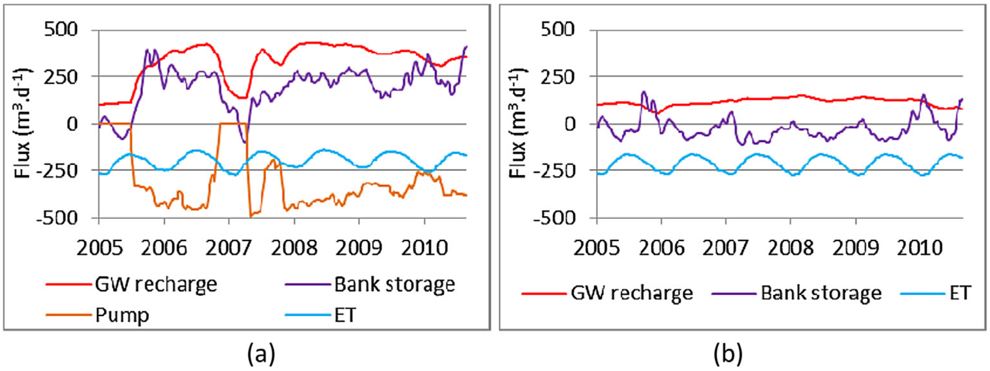

Fig. 7. Groundwater balance for Only-SIS (a) and No-salt management scenario (b).

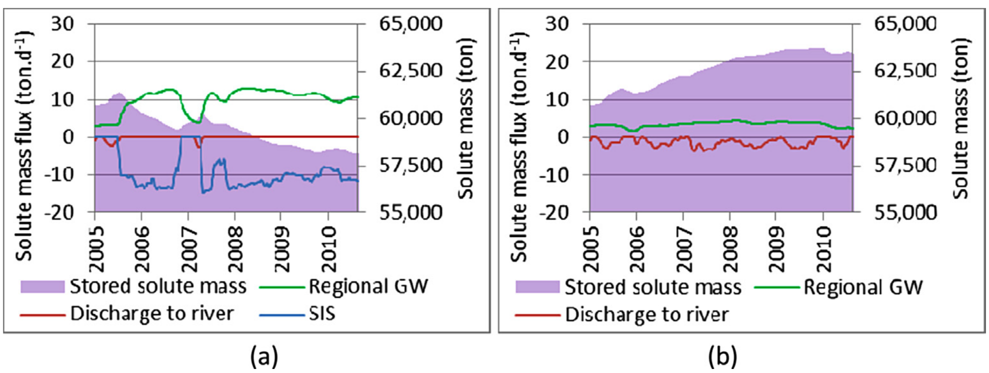

Fig. 8. Solute mass balance for Only-SIS scenario (a) and No-salt management scenario (b).

discharged to the river after the trial. This is partly due to bank storage and also to a loss of water to evapotranspiration since each trial provides more water available for evapotranspiration. For example, as more water enters the floodplain aquifer, consequently more water returns back to the river during the river stage recession and this results in a greater flux for the 1.5 m scenario than for the other scenarios.

During the river drop scenario a different behaviour can be observed. In that period the floodplain is in a losing condition and the hydraulic gradient is changed towards the river. But, after each trial, there is a lag-time before the system reaches an equilibrium. During this lag-time, a significant amount of water moves from the river to the floodplain aquifer. The time required for the system to reach the equilibrium appears to be more closely related to the duration of the trial rather than to the river stage level changes.

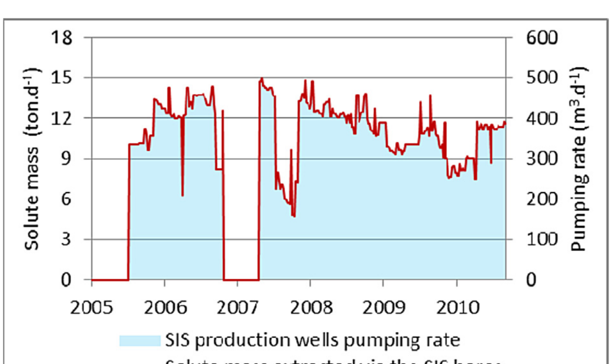

Fig. 9. Solute mass extracted via the SIS production wells during the study period.

It seems the general pattern of flux exchange is similar for all four trials and the SIS operation does not have a significant influence on the pattern. Even, the slight increase in bank storage during the 1st round of SIS operation (July 2005–November 2006) strongly corresponds to the river stage fluctuation. In other words, river stage manipulation has more influence than SIS operation on the groundwater head.

Another important component of the water balance is floodplain aquifer recharge from regional groundwater, which is shown in Fig. 12c. Recharge from regional groundwater is strongly attributed to the operation of the SIS production wells. In fact, recharge increases by three to four-fold during SIS operation compared to either before their operation (before July 2005) or when the SIS wells were shut down (from November 2006 to April 2007). However, a higher river stage rise may decrease the flux from the regional groundwater to the floodplain aquifer due to the enhanced hydraulic head towards the floodplain aquifer during the river stage rise trials. For example, the 0.5 m drop scenario leads to a slight increase in regional groundwater recharge.

3.3. Solute mass balance

In order to analyse the spatial and temporal solute dynamics of the floodplain aquifer in the context of SW–GW interaction, the following modelled outputs were considered: (a) total stored solute mass in the floodplain aquifer; (b) change in total stored solute mass in the floodplain aquifer; (c) stored solute mass in the unsaturated zone; and (d) solute concentrations at observation wells B1 and B4.

The total stored solute mass in the floodplain aquifer is shown in Fig. 13a. Decreases in stored solute mass in the floodplain aquifer

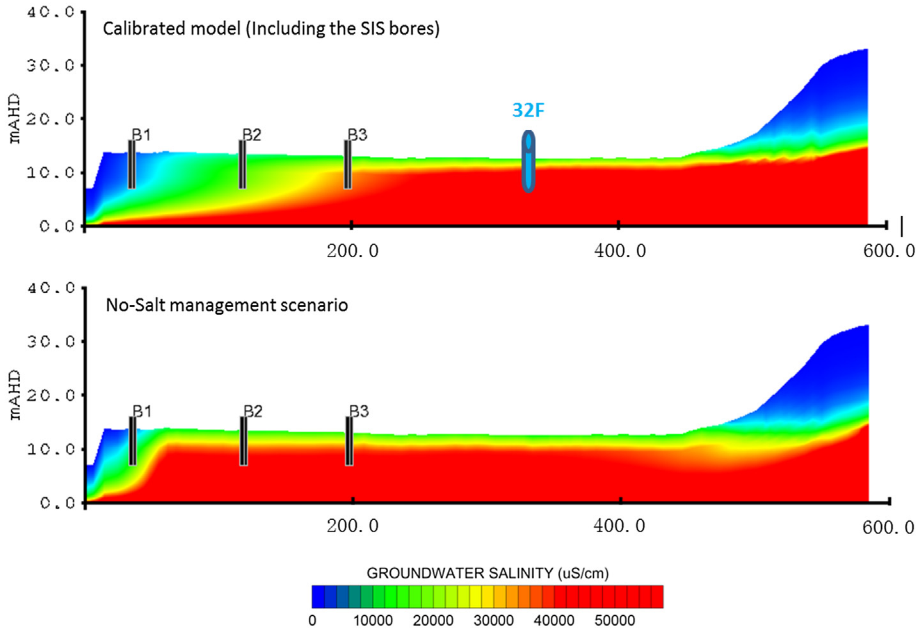

Fig. 10. Groundwater salinity along transect 1 for Only-SIS and No-salt management scenarios (Z magnification: 3) at time step 2070 days (2/09/2010).

are mainly attributed to saline groundwater extraction by the SIS production wells. This is why the total stored solute mass generally decreases for all scenarios. However, it may be seen that a higher river stage results in relatively less solute being accumulated in the floodplain aquifer. Since the main source of solute mass entering the floodplain aquifer is regional saline groundwater, an increase in river stage can reduce the rate of saline groundwater entering the floodplain aquifer. On the other hand, the 0.5 m drop scenario results in relatively more solute accumulation. This is due to the increased recharge of the floodplain aquifer from saline regional groundwater, which is consistent with the results shown in Fig. 12c.

The results indicate that for a higher stage rise, comparatively less solute mass accumulates in the floodplain aquifer. This is because of the increase in head gradient from the river towards the floodplain aquifer. Fig. 13b illustrates the change in stored solute mass in the floodplain aquifer over the study period. As a general

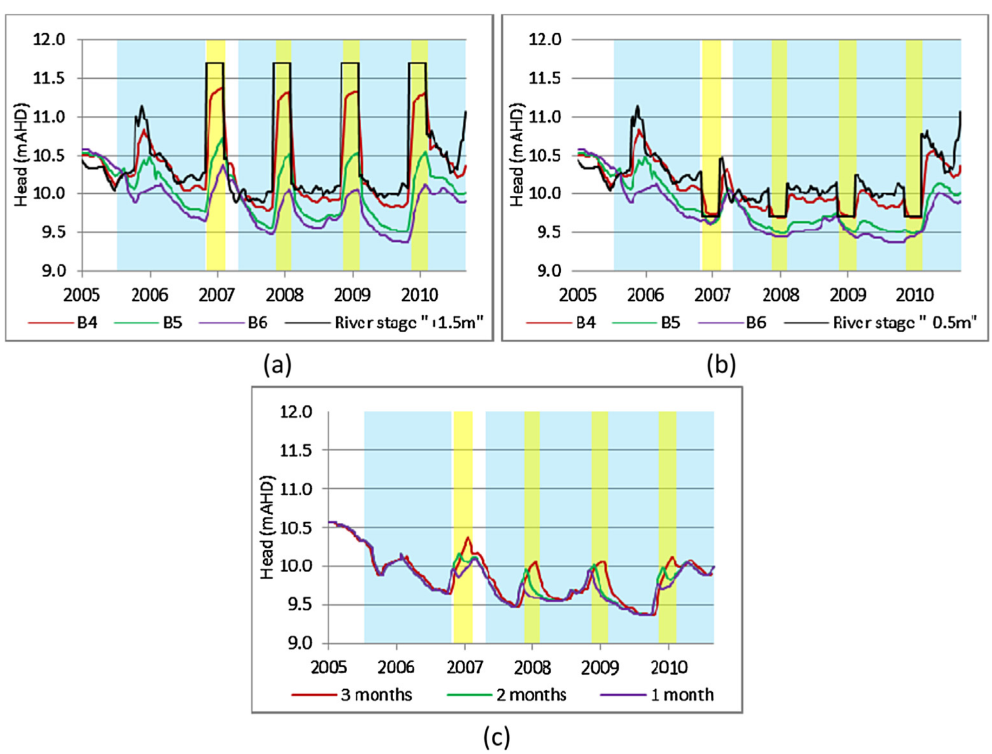

Fig. 11. Dynamics of GW heads at the observation wells on Transect 2, a: 1.5 m rise for 3 months, b: 0.5 m drop for 3 months and c: 1.5 m rise for one, two and three months at observation well B6.

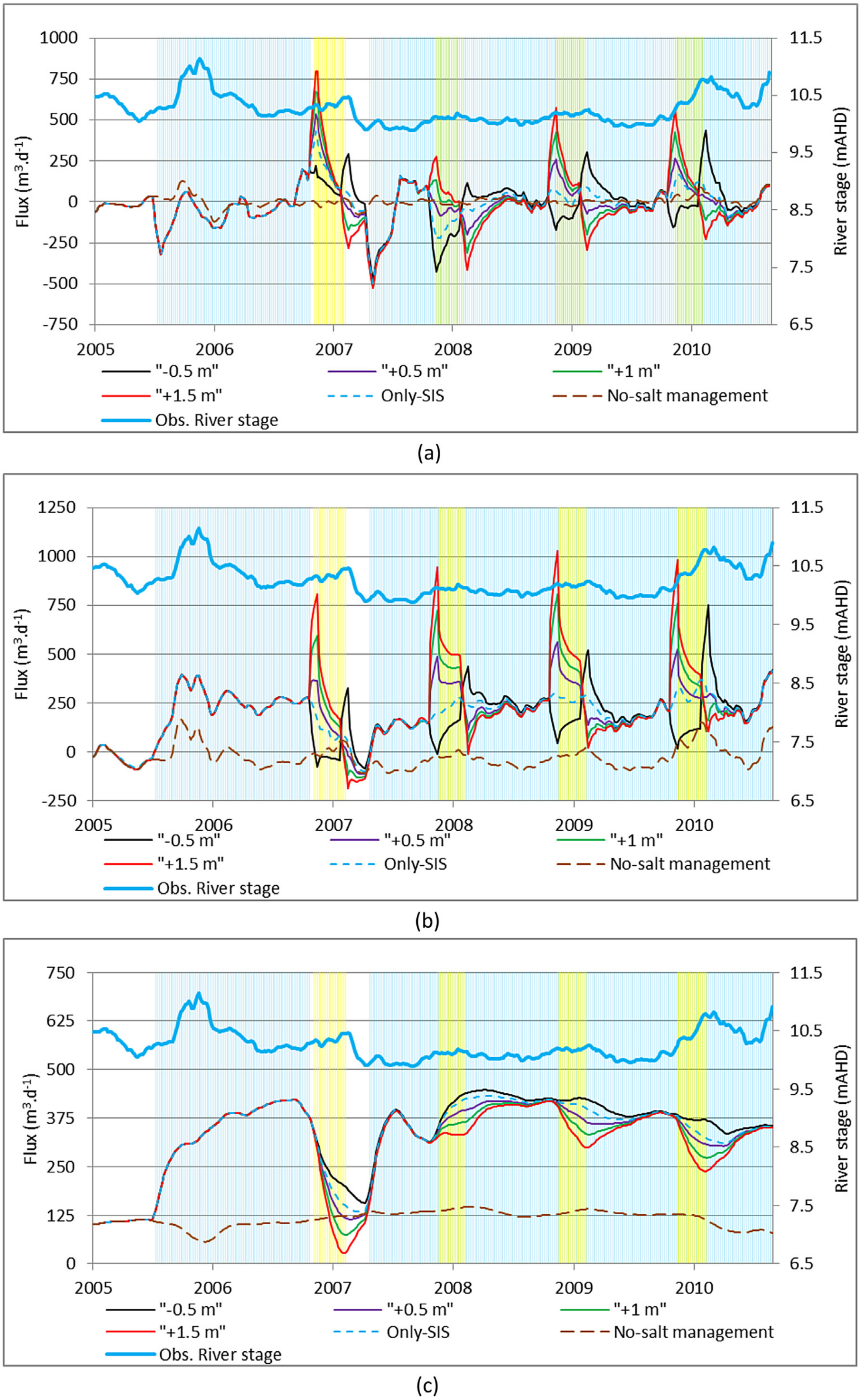

Fig. 12. a: Change in water storage in the floodplain aquifer, b: Flux exchange between the river and the floodplain aquifer and c: Floodplain aquifer recharge from regional groundwater. All the results shown here are for the three month scenarios. The blue and yellow patterns represent groundwater lowering and river stage manipulation, respectively. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

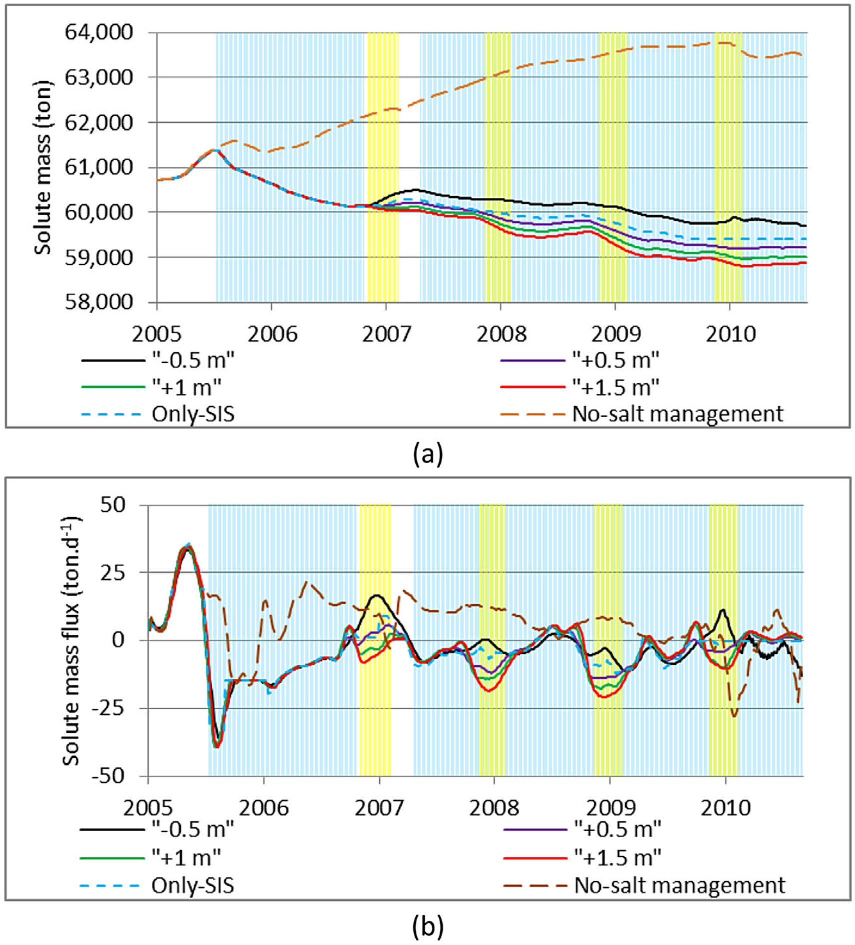

Fig. 13. a: Total solute mass in the floodplain aquifer and b: Change in stored solute mass in the floodplain aquifer. Both are for the three month scenarios. The blue and yellow patterns represent groundwater lowering and river stage manipulation, respectively. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

pattern, when the SIS operates, the change in stored solute mass is negative (more solute leaves the system than enters). Higher river stage rises may enhance this process by increasing the hydraulic gradient towards the floodplain aquifer. In contrast, when the floodplain aquifer is losing, it means that more saline groundwater from the regional aquifer enters the floodplain. This can be seen in the 0.5 m drop scenario.

Changes in solute concentration along transect 1 due to changes in river stage for the 0.5 m drop and +1.5 m rise three month scenarios during the 3rd trial are presented in Fig. 14. Clearly, a 0.5 m drop scenario produces a more saline aquifer, while the 1.5 m rise would produce the least saline aquifer. Also, the extent of the freshwater lens is longer in the 1.5 m rise scenario. Considering Figs. 11 and 14, it appears that the river stage manipulation operation is more effective in the vicinity of the river bank. This can be beneficial in arid and semi-arid floodplains where riparian vegetation health depends on the availability of freshwater, as is the case in the study area. It is also expected that larger increases in river stage will result in (a) dilution propagating further inland and (b) dilution being more pronounced at the river-aquifer interface.

3.4. Solute mass in the unsaturated zone

It was found that at the beginning of the study period, 13% of the total solute mass was stored in the unsaturated zone. Two main drivers are influencing the solute dynamic in the floodplain aquifer including groundwater lowering and river stage manipulation. Here, the impact of each of these drivers on the accumulated or mobilized solute mass in the unsaturated zone is analysed.

Fig. 15 compares the solute mass in the unsaturated zone in No-salt management and Only-SIS along scenarios along transect 2. As expected, a significant amount of solute mass is mobilized due to the operation of the SIS production wells. This may be because the SIS operation lowers the groundwater table which leads to an overall less saline unsaturated zone. Moreover, Only-SIS scenario shows a much larger freshwater lens compared to No-salt management scenario. Again, it seems that groundwater lowering is able to maintain a less saline floodplain aquifer by mitigating regional groundwater recharge on one hand and by attracting more freshwater via the river bank on the other hand. Furthermore, the results confirm that Only-SIS scenario at the last time-step (2070) shows a mobilization of 6% of the solute mass from the unsaturated zone. On the other hand, No-salt management scenario shows 5% more solute mass accumulation in the unsaturated zone compared to the model in the 1st time step.

Fig. 16 shows the spatial distribution of the solute concentration in the floodplain aquifer for the three month long 1.5 m rise and 0.5 m drop scenarios at time-step 1120 days (just after the 2nd trial) along transect 2. Comparing with Fig. 15, it appears that river stage manipulation has less influence than groundwater lowering floodplain salinity, with the influence being limited to the near-river zone. The higher hydraulic gradient from the river to the floodplain creates

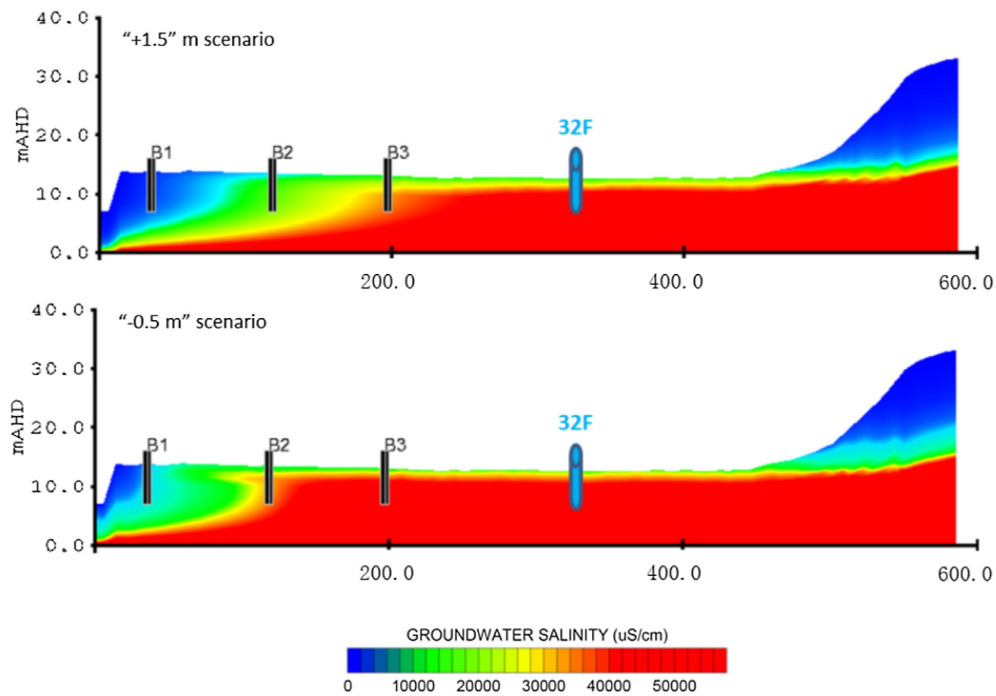

Fig. 14. Spatial distribution of modelled solute concentration along transect 1 during the 3rd trial (time step 1480 (20/01/2009)) for the 1.5 m rise and 0.5 m drop three month scenarios. Observation wells are shown in black.

a relatively larger freshwater lens. The results indicate that the total solute mass in the unsaturated zone under the 1.5 m rise scenario is 7% less than for the 0.5 m drop scenario after the 2nd trial. In fact, the 0.5 m drop scenario leads to more saline regional groundwater recharges to the floodplain (see Fig. 13b).

In order to demonstrate the impact of river stage manipulation on the solute mass accumulation in the unsaturated zone, time-steps 0 (beginning of the study period) and 1120 days (after the 2nd trial) in the three month 1.5 m rise scenario are analysed. Fig. 17 shows the spatial distribution of solute mass mobilization attributed to river stage manipulation (three month, +1.5 m scenario) at time-step 1120 days. The model results show that up to 14 kg m^{-3} of solute mass is removed from the unsaturated zone. It appears that the greatest solute mass mobilization occurs at a distance of around 50–100 m from the river. In fact, the extent that river stage manipulation may significantly affect the solute mobilization is limited to the extent that it could change the head gradient, which in this case is the boundary of the fresh and saline zones. The solute mass mobilization in the rest of the floodplain aquifer is restricted. This is because of the relatively lower amount of solute stored at the river bank (0–50 m). On the other hand, as the main source of solute comes from the highland aquifer, less change in solute in the unsaturated zone can be expected in the highland.

3.5. Ecological implications

Knowledge of the interaction between groundwater and surface water bodies is vital for assessing the role of riparian floodplain processes on water quality and groundwater level dynamics (Eslamian and Nekoueineghad, 2009; Rassam, 2005). The ecological implications of river stage manipulation were not the foremost objective of this research. However, it can be inferred that each river stage rise trial may lead to soil water freshening and this may lead to a riparian tree response since each trial makes more fresh water accessible for riparian trees. This is reinforced by the results of Berens et al. (2009), Holland et al. (2009a, 2013) who showed that introduction of sufficient fresh water through sources such as artificial inundation and bank storage to a saline floodplain aquifer can to some extent maintain species richness and diversity on the floodplain. For instance, the availability of relatively more fresh water (through a freshened soil profile) may lead to epicormic growth in trees. However, the ecological response mainly depends on the vegetation condition at the beginning of the trial (Holland et al., 2013). Therefore, due to the overall saline nature of arid and semi-arid floodplains, this response is unlikely to be sustained unless there is a regular recurrence of the trial cycles.

4. Conclusion

A fully integrated, physically-based, numerical model (HydroGeoSphere) of surface water–groundwater flow and solute transport at Clark's Floodplain was developed and calibrated against observed data, which included river stages, floodplain aquifer heads and solute concentrations. The calibration results showed that the model was capable of reproducing the dynamics of both the flow and the solute. The calibrated model was applied to investigate the relative impacts of river stage manipulation on surface water–groundwater interactions that were also being influenced by Salt Interception Scheme (SIS) groundwater lowering measures. Twelve hypothetical scenarios were defined including river stage changes of +0.5, +1.0, +1.5, and −0.5 m.

Using the calibrated model (Only-SIS scenario), it was shown that groundwater lowering via the SIS production wells was able to mitigate the saline regional groundwater recharge on the one hand and enhance river bank storage on the other hand, thereby leading to a less saline floodplain aquifer. In fact, by lowering the groundwater table, the SIS production wells created a divide which stopped saline water from reaching the floodplain. In this situation, even if groundwater discharge to the river via the river bank did occur, it would not have a significant impact on the river water quality. However, without the groundwater lowering operation, there would be no process to prevent the saline groundwater from reaching the river and the only discharge process would be via the river banks. It was demonstrated that an absence of groundwater

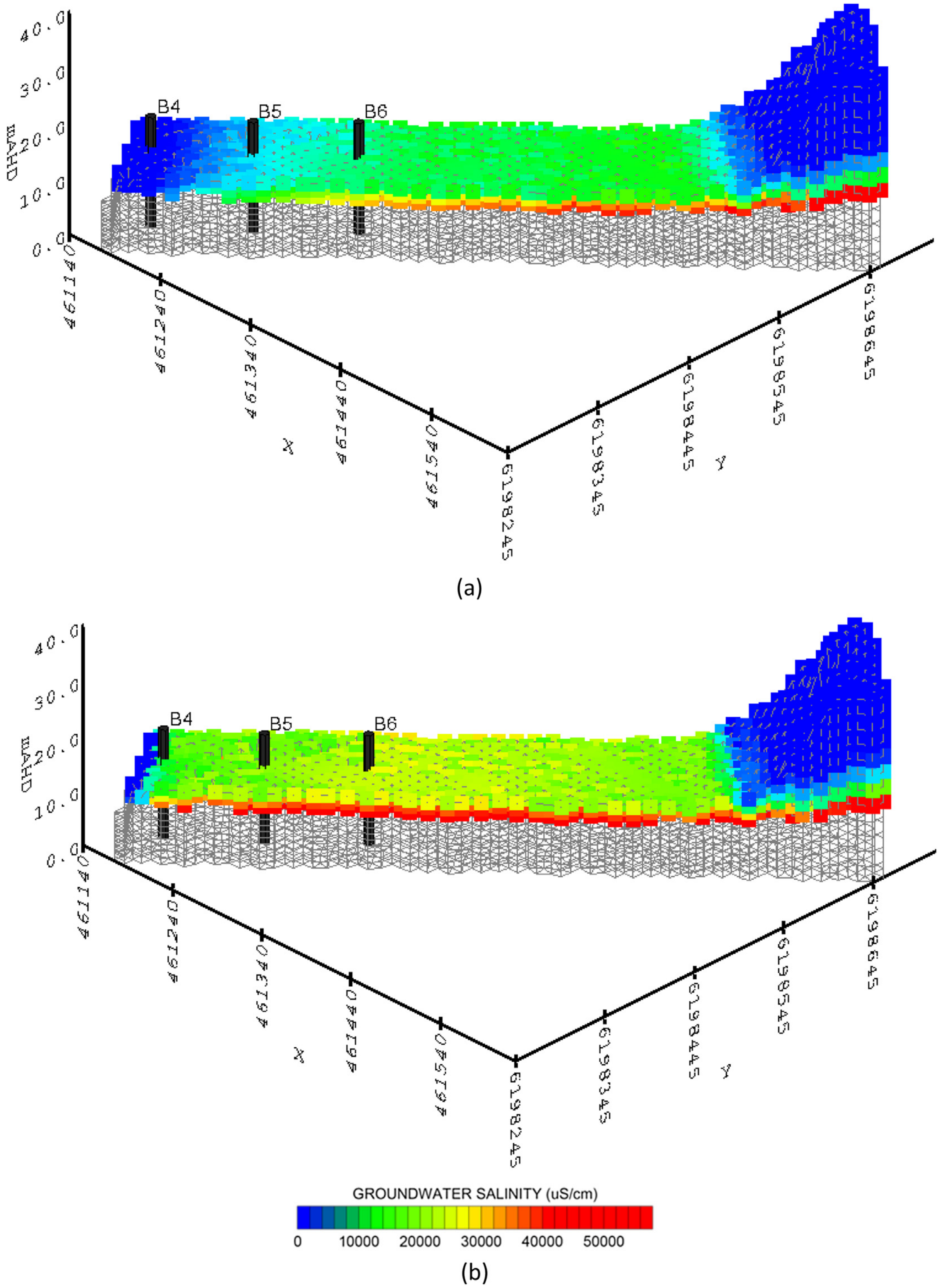

Fig. 15. Visualization of the solute concentration distribution in the floodplain aquifer for Only-SIS (a) and No-salt management scenarios (b) along transect 2.

lowering measures would have led to continuous saline groundwater discharge to the river which could be a risk to the river water quality over the long term. In other words, the SIS was successful in intercepting saline groundwater that would otherwise have entered the river.

The hypothetical scenarios were used to demonstrate the impact of river stage manipulation on the SW–GW interaction in the river-floodplain system. In terms of water balance, it was shown that each rising river manipulation scenario led to a gaining floodplain which was proportional to the height and duration of each trial. The floodplain aquifer became a losing feature after each rising trial for a duration almost equal to the duration of the trial. This pattern was reversed for the 0.5 m drop scenario. It was shown that river stage rise formed higher gradients from the river to the floodplain aquifer and this was much more pronounced in the near-river zone (up to 100 m from the river). However, longer durations of river stage manipulation could potentially extend the influence further into the floodplain.

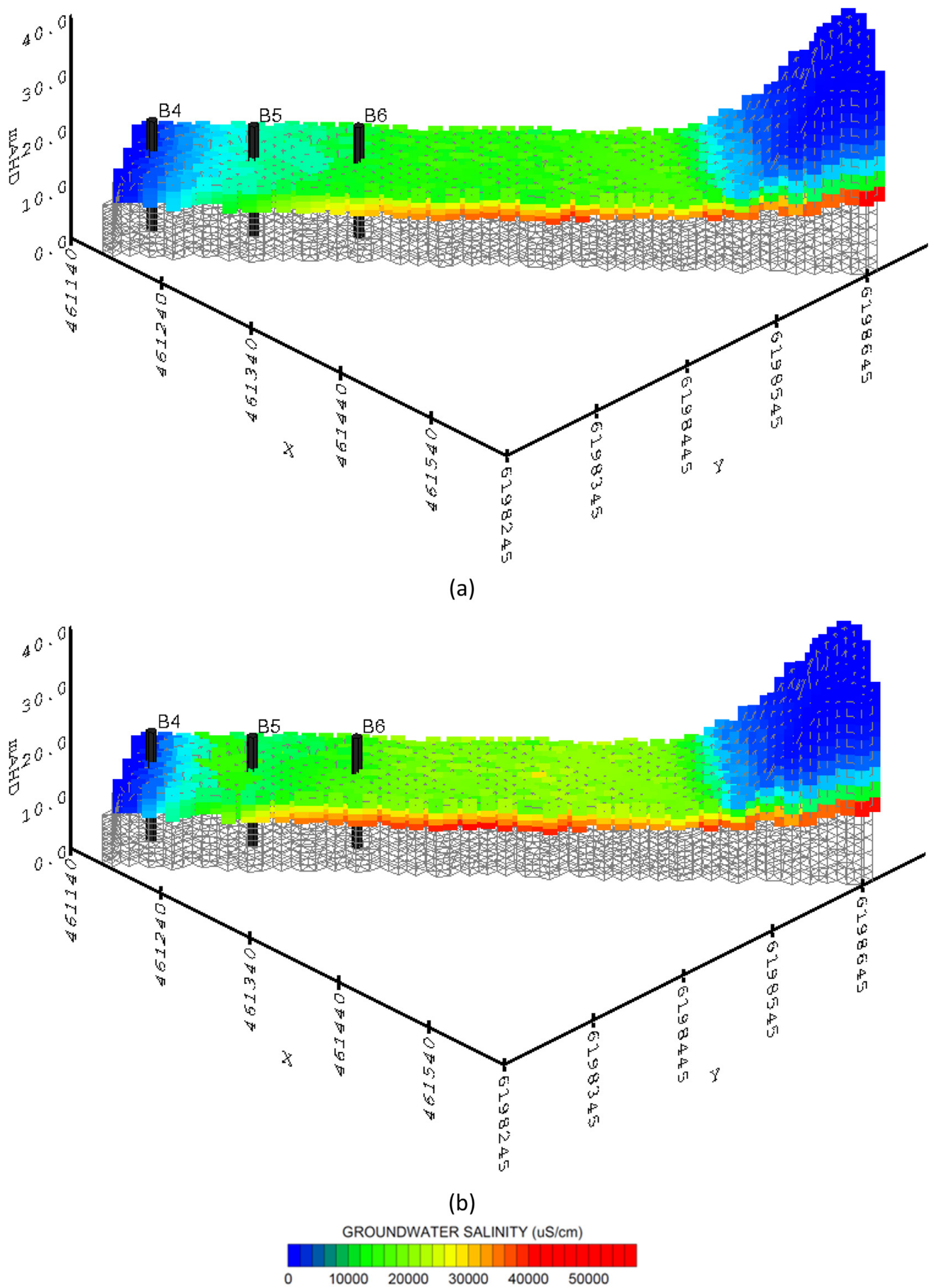

Fig. 16. Visualization of the solute concentration distribution in the floodplain aquifer for the three month long +1.5 m (a) and −0.5 m (b) scenarios at time-step 1120 days (just after the 2nd trial) along transect 2.

Furthermore, it was demonstrated that a higher river stage rise was able to reduce the flux from the regional groundwater to the floodplain aquifer due to the enhanced hydraulic head towards the floodplain aquifer. Consequently, a higher river stage led to relatively less solute mass in the floodplain aquifer. In contrast, scenarios that involve lowering the river stage enhanced the flux from the saline regional groundwater to the floodplain aquifer. This resulted in a more saline floodplain aquifer.

Among the dominant drivers in this study, groundwater extraction had a greater influence on solute mass mobilization in the unsaturated zone. However, river stage rise can also be effective but is limited to the near-river zone (up to 100 m). Moreover, the results of this study indicate that river stage manipulation may be able to decrease the soil and groundwater salinities in the near-river zone which can improve the health of riparian vegetation. However, its impacts are spatially and temporally limited and river

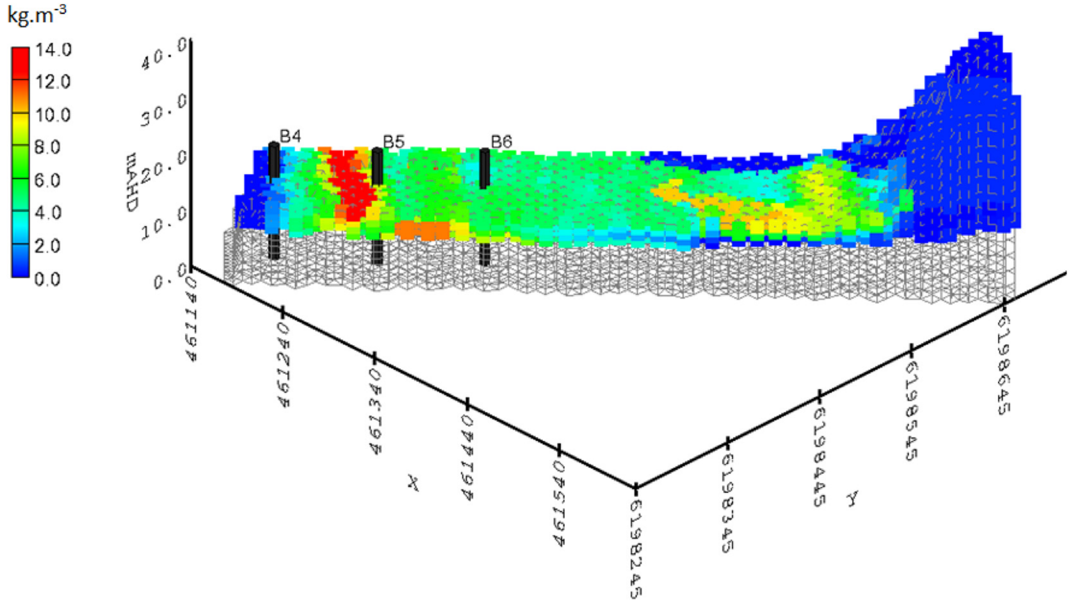

Fig. 17. Visualization of distribution of solute mass mobilization in the floodplain aquifer for the three month long, +1.5 m scenario at time-step 1120 days (just after the 2nd trial) along transect 2.

stage manipulation cannot entirely change the natural condition of the floodplain. Hence, it may be considered as a short term management technique. However, if longer term strategies are required, it may be possible to implement these salt interception measures periodically.

According to the results of this study, it can be concluded that bank storage is one of the main drivers of surface–groundwater interactions in a saline, semi-arid floodplain, particularly in non-flooding conditions. Also, induced bank storage through river level manipulation may lead to a less saline floodplain aquifer. This happens through two mechanisms. First, an increase in river stage lowers the gradient from the regional groundwater aquifer to the floodplain which reduces saline groundwater recharge to the floodplain. Second, modelling supports observational data that bank storage is able to freshen the groundwater near the river banks during high-flow pulses by mixing fresh water with saline groundwater. The applicability of these findings to other areas depends on floodplain topography, soil salinity, groundwater condition (i.e. gaining or losing), geomorphology of the floodplain aquifer (i.e. hydraulic conductivity), river characteristics (i.e. bed conductance, bank slope, depth) and vegetation condition. Further studies are recommended to investigate other potential drivers and salt interception measures, such as artificial floodplain inundation.

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References

Alaghmand, S., Beecham, S., Hassanli, A., 2013a. Impacts of groundwater extraction on salinization risk in a semi-arid floodplain. Nat. Hazards Earth Syst. Sci. 13 (12), 3405–3418.
Alaghmand, S., Beecham, S., Hassanli, A., 2013b. A review of the numerical modelling of salt mobilization from groundwater-surface water interactions. Water Resour. 40 (3), 325–341.
Allison, G.B., Cook, P.G., Barnett, S.R., Walker, G.R., Jolly, I.D., Hughes, M.W., 1990. Land clearance and river salinization in the western Murray Basin, Australia. J. Hydrol. 119, 1–20.
AquaVeo, 2011. GMS: Provo, UT.
AWE, 1999. Clarks Floodplain Investigations. Prepared for the Loxton to Bookpurnong Local Action Planning Committee. Australian Water Environments Report Adelaide.
AWE, 2013. Loxton Bookpurnong SIS Atlas. Australian Water Environment, Adelaide.
Banks, E.W., Brunner, P., Simmons, C.T., 2011. Vegetation controls on variably saturated processes between surface water and groundwater and their impact on the state of connection. Water Resour. Res. 47 (11).
Barnett, B., Townley, L.R., Post, V., Evans, R.E., Hunt, R.J., Peeters, L., Richardson, S., Werner, A.D., Knapton, A., Boronkay, A., 2012. Australian groundwater modelling guidelines. In: report, W. (Ed.). National Water Commission, Canberra.
Barnett, S.R., Yan, Y., Watkins, N.R., Woods, J.A., Hyde, K.M., 2002. Murray Darling Basin Salinity Audit: Groundwater Modelling to Predict Future Salt Loads to the River Murray in South Australia. Report DWR 2001/017. Department for Water Resources, Adelaide.
Bauer, P., Gumbricht, T., Kinzelbach, W., 2006a. A regional coupled surface water/groundwater model of the Okavango Delta, Botswana. Water Resour. Res. 42 (4). W04403.
Bauer, P., Held, R.J., Zimmermann, S., Linn, F., Kinzelbach, W., 2006b. Coupled flow and salinity transport modelling in semi-arid environments: the Shashe River Valley, Botswana. J. Hydrol. 316 (1–4), 163–183.
Bennett, N.D., Croke, B.F.W., Guariso, G., Guillaume, J.H.A., Hamilton, S.H., Jakeman, A.J., Marsili-Libelli, S., Newham, L.T.H., Norton, J.P., Perrin, C., Pierce, S.A., Robson, B., Seppelt, R., Voinov, A.A., Fath, B.D., Andreassian, V., 2013. Characterising performance of environmental models. Environ. Model. Softw. 40, 1–20.
Berens, V., White, M., Souter, N., 2009. Bookpurnong Living Murray Pilot Project: a Trial of three floodplain Water Management Techniques to Improve Vegetation Condition. Department of Water, Land and Biodiversity Conservation, Adelaide.
BOM, 2013. Bureau of Meteorology (BOM).
Brooks, R.J., Corey, A.T., 1964. Hydraulic Properties of Porous Media. Hydrology Paper No. 3. Colorado State University, Fort Collins.
Brunner, P., Cook, P.G., Simmons, C.T., 2009. Hydrogeologic controls on disconnection between surface water and groundwater. Water Resour. Res. 45 (1). W01422.
Brunner, P., Simmons, C.T., 2012. HydroGeoSphere: a fully integrated, physically based hydrological model. Ground Water 50 (2), 170–176.
Carsel, R.F., Parrish, R.S., 1988. Developing joint probability distributions of soil water retention characteristics. Water Resour. Res. 24 (5), 755–769.
Crowe, A.S., Shikaze, S.G., Ptacek, C.J., 2004. Numerical modelling of groundwater flow and contaminant transport to Point Pelee marsh, Ontario, Canada. Hydrol. Process. 18 (2), 293–314.
Doble, R., Brunner, P., McCallum, J., Cook, P.G., 2012. An analysis of river bank slope and unsaturated flow effects on bank storage. Ground Water 50 (1), 77–86.

Doble, R., Simmons, C., Jolly, I., Walker, G., 2006. Spatial relationships between vegetation cover and irrigation-induced groundwater discharge on a semi-arid floodplain, Australia. J. Hydrol. 329 (1–2), 75–97.
Ehret, U., Zehe, E., 2011. Series distance – an intuitive metric to quantify hydrograph similarity in terms of occurrence, amplitude and timing of hydrological events. Hydrol. Earth Syst. Sci. 15 (3), 877–896.
Eslamian, S., Nekoueineghad, B., 2009. A review on interaction of groundwater and surface water. Int. J. Water 5 (2), 89–99.
Evans, R., Hoxley, G., Collett, K., 2013. Mallee Salinity Workshop: Chapter 2-Floodplain Processes. Mallee Catchment Management Authority.
Ewen, J., 2011. Hydrograph matching method for measuring model performance. J. Hydrol. 408 (1–2), 178–187.
Ghazavi, R., Vali, A.B., Eslamian, S., 2012. Impact of flood spreading on groundwater level variation and groundwater quality in an arid environment. Water Resour. Manag. 26 (6), 1651–1663.
Gupta, H.V., Clark, M.P., Vrugt, J.A., Abramowitz, G., Ye, M., 2012. Towards a comprehensive assessment of model structural adequacy. Water Resour. Res. 48 (8).
Hart, B., Bailey, P., Edwards, R., Hortle, K., James, K., McMahon, A., Meredith, C., Swadling, K., 1991. A review of the salt sensitivity of the Australian freshwater biota. Hydrobiologia 210 (1), 105–144.
Herczeg, A.L., Simpson, H.J., Mazor, E., 1993. Transport of soluble salts in a large semiarid basin: river Murray, Australia. J. Hydrol. 144, 59–84.
Hingston, F.J., Galbraith, J.H., Dimmock, G.M., 1997. Application of the process-based model BIOMASS to Eucalyptus globules subsp. Globules plantations on ex-farmland in south Western Australia: I. Water use by trees and assessing risk of losses due to drought. For. Ecol. Manag. 106, 141–156.
Hoehn, E., Scholtis, A., 2011. Exchange between a river and groundwater, assessed with hydrochemical data. Hydrol. Earth Syst. Sci. 15 (3), 983–988.
Holland, K.L., Charles, A.H., Jolly, I.D., Overton, I.C., Gehrig, S., Simmons, C.T., 2009a. Effectiveness of artificial watering of a semi-arid saline wetland for managing riparian vegetation health. Hydrol. Process. 23, 3474–3484.
Holland, K.L., Jolly, I.D., Overton, I.C., Walker, G.R., 2009b. Analytical model of salinity risk from groundwater discharge in semi-arid, lowland floodplains. Hydrol. Process. 23, 3428–3439.
Holland, K.L., Turnadge, C.J., Nicol, J.M., Gehrig, S.L., Strawbridge, A.D., 2013. Floodplain Response and Recovery: Comparison Between Natural and Artificial Floods. Technical Report Series No. 13/4. Goyder Institute for Water Research, Adelaide.
Jakeman, A.J., Letcher, R.A., Norton, J.P., 2006. Ten iterative steps in development and evaluation of environmental models. Environ. Model. Softw. 21 (5), 602–614.
Jolly, I.D., 2004. Review of River Murray Floodplain Salinity Studies in South Australia and Their Relevance to the Victorian Mallee Floodplains. CSIRO Land and Water, Adelaide.
Jolly, I.D., McEwan, K.L., Holland, K.L., 2008. A review of groundwater-surface water interactions in arid/semi-arid wetlands and the consequences of salinity for wetland ecology. Ecohydrology 1 (1), 43–58.
Jolly, I.D., Walker, G.R., Hollingworth, I.D., Eldridge, S.R., Thorburn, P.J., McEwan, K.L., Hatton, T.J., 1996. The causes of decline in eucalypt communities and possible ameliorative approaches. In: walker, G.R., Jolly, I.D., Jarwal, S.D. (Eds.), Salt and Water Movement in the Chowilla Floodplain. CSIRO Division of Water Resources, Canberra, Australia.
Jolly, I.D., Walker, G.R., Thorburn, P.J., 1993. Salt accumulation in semi-arid floodplain soils with implications for forest health. J. Hydrol. 150 (2–4), 589–614.
Knight, J.H., Rassam, D.W., 2007. Groundwater head responses due to random stream stage fluctuations using basis splines. Water Resour. Res. 43 (6).
Kollet, S.J., Maxwell, R.M., Woodward, C.S., Smith, S., Vanderborght, J., Vereecken, H., Simmer, C., 2010. Proof of concept of regional scale hydrologic simulations at hydrologic resolution utilizing massively parallel computer resources. Water Resour. Res. 46 (4).
Langevin, C., Swain, E., Wolfert, M., 2005. Simulation of integrated surface-water/ground-water flow and salinity for a coastal wetland and adjacent estuary. J. Hydrol. 314 (1–4), 212–234.
Matthews, K.B., Rivington, M., Blackstock, K., McCrum, G., Buchan, K., Miller, D.G., 2011. Raising the bar? – the challenges of evaluating the outcomes of environmental modelling and software. Environ. Model. Softw. 26 (3), 247–257.
McCallum, J.L., Cook, P.G., Brunner, P., Berhane, D., 2010. Solute dynamics during bank storage flows and implications for chemical base flow separation. Water Resour. Res. 46 (7).
McEwan, K.L., Jolly, I.D., Holland, K.L., 2006. Groundwater-Surface Water Interactions in Arid/Semi-Arid Wetlands and the Consequences of Salinity for Wetland Ecology. Science Report 53/06. CSIRO Land and Water, Adelaide.
McLaren, R.G., 2005. Grid Builder: A Pre-processor for 2-D, Triangular Element, Finite-element Programs. Groundwater Simulations Group, University of Waterloo, Waterloo, Ontario.
Meire, D., De Doncker, L., Declercq, F., Buis, K., Troch, P., Verhoeven, R., 2010. Modelling river-floodplain interaction during flood propagation. Nat. Hazards 55 (1), 111–121.
Nash, J.E., Sutcliffe, J.V., 1970. River flow forecasting through conceptual models part I – a discussion of principles. J. Hydrol. 10 (3), 282–290.
Peck, A.J., Hatton, T., 2003. Salinity and the discharge of salts from catchments in Australia. J. Hydrol. 272, 191–202.
Peck, A.J., Hurle, D.H., 1973. Chloride balance of some farmed and forested catchments in Southwestern Australia. Water Resour. Res. 9, 648–657.
Rassam, D.W., 2005. Impacts of Hillslope Floodplain Characteristics on Groundwater Dynamics: Implications for Riparian Denitrification. the International Congress on Modelling and Simulation, Melbourne, Australia, pp. 2735–2741.
Rassam, D.W., 2011. A conceptual framework for incorporating surface–groundwater interactions into a river operation–planning model. Environ. Model. Softw. 26 (12), 1554–1567.
Rassam, D.W., Peeters, L., Pickett, T., Jolly, I., Holz, L., 2013. Accounting for surface-groundwater interactions and their uncertainty in river and groundwater models: a case study in the Namoi River, Australia. Environ. Model. Softw. 50, 108–119.
Restrepo, J.I., Montoya, A.M., Obeysekera, J., 1998. A wetland simulation Module for the MODFLOW ground water model. Ground Water 36 (5), 764–770.
Schmid, W., Hanson III, R.T., Thomas, M.M., Leake, S.A., 2006. User's guide for the Farm process (FMP) for the U.S. Geological Survey's Modular Three-dimensional Finite Difference Ground-water Flow Model, MODFLOW-2000, USGS Techniques and Methods 6-A17. USGS, Reston, Virginia.
Sophocleous, M., 2010. Revue critique: Pratiques, défis et innovations dans le domaine des de la gestion des eaux souterraines de l'aquifère des Grandes Plaines (High Plains), aux Etats Unis d'Amérique – Leçons et recommandations (Review: groundwater management practices, challenges, and innovations in the High Plains aquifer, USA-lessons and recommended actions). Hydrogeol. J. 18 (3), 559–575.
Straatsma, M.W., van der Perk, M., Schipper, A.M., de Nooij, R.J.W., Leuven, R.S.E.W., Huthoff, F., Middelkoop, H., 2013. Uncertainty in hydromorphological and ecological modelling of lowland river floodplains resulting from land cover classification errors. Environ. Model. Softw. 42, 17–29.
Telfer, A., Overton, I.C., 1999. he impact of irrigation on floodplain vegetation health adjacent the River Murray. In: Rutherford, I.D., Bartley, R. (Eds.), T. The Second Australian Stream Management Conference, Adelaide.
Therrien, R., McLaren, R.G., Sudicky, E.A., Panday, S.M., 2006. HydroGeoSphere: A Three-Dimensional Numerical Model Describing Fully-Integrated Subsurface and Surface Flow and Solute Transport. Groundwater Simulations Group, University of Waterloo, Waterloo, Canada.
van Genuchten, M.T., 1980. A closed-form equation for predicting the hydraulic conductivity of unsaturated soils. Sci. Soc. Am. J. 44 (5), 892–898.
Verstrepen, L., 2011. Evaluating Rainwater Harvesting on Watershed Level in the Semi-arid Zone of CHILE, Bioscience Engineering. Universiteit Gent, Gent, p. 113.
Walton, R., Chapman, R., Davis, J., 1996. Development and application of the wetlands Dynamic Water Budget Model. Wetlands 16 (3), 347–357.
WaterConnect, 2013. River Murray Water Data. https://www.waterconnect.sa.gov.au/RMWD/Pages/default.aspx.
White, M.G., Berens, V., Souter, N.J., 2009. Bookpurnong Living Murray Pilot Project: Artificial Inundation of *Eucalyptus camaldulensis* on a Floodplain to Improve Vegetation Condition. Science, Monitoring and Information Division, Department of Water, Land and Biodiversity Conservation.
Zimmermann, S., Bauer, P., Held, R., Kinzelbach, W., Walther, J.H., 2006. Salt transport on islands in the Okavango Delta: numerical investigations. Adv. Water Resour. 29 (1), 11–29.