Modelling salt dynamics on the River Murray floodplain in South Australia: Modelling approaches

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Executive summary

The River Murray is the principal river of the western Murray Basin in southeastern Australia. It is highly unusual as it is a major river system that flows through an extensive landscape with highly saline groundwater. The river is naturally prone to salinity, and this propensity has increased over the past century due to the construction of river locks and the introduction of large-scale land clearance and irrigation. River flow volumes and river level variability have been reduced, while the watertable has risen. The overall impact has been to degrade riverine ecosystems and increase river and floodplain salinity.

Management of the lower River Murray floodplain requires a detailed understanding of its hydrological and hydrogeological processes, including those which mobilise or store salt. The Basin Plan increases State obligations to manage and report on salinity and water quality targets for the River Murray. A major outcome of the Basin Plan is to deliver environmental flows to help protect and restore River Murray wetlands and floodplains. To do so effectively, SA must understand the short-term movement of water and salt within the floodplain landscape, under present conditions and under various management options for delivering environmental water.

Much research has been undertaken to monitor, conceptualise and simulate the dynamics of floodplain salinity. However, there is currently no consistent and comprehensive approach to modelling the interaction of surface water and groundwater in the lower River Murray floodplains. To address this need, in 2014 the Goyder Institute for Water Research commissioned the study *Modelling salt dynamics on the River Murray floodplain in South Australia*, a collaborative research project with contributions from Flinders University, CSIRO Land and Water, and the Department of Environment, Water and Natural Resources (DEWNR). The study area consists of the floodplains of the River Murray in South Australia, from the Border to Morgan, within the South Australian part of the Murray Basin. Due to the breadth of the review required, feedback was sought from a wide range of experts and stakeholders.

The companion report, *Modelling salt dynamics on the River Murray floodplain in South Australia: Conceptual model, data review and salinity risk approaches* (Woods, 2015a), documents the resulting conceptual model, data review, and salinity risk methodology discussion. This report documents the literature review and testing of modelling approaches. A third report contains the Appendices (Woods, 2015b).

Review of current approaches to groundwater modelling

Physically-based numerical groundwater models of floodplains or floodplain processes in Australia have been developed predominantly with the industry standard groundwater model MODFLOW and in some cases its solute transport counterpart MT3DMS. Other models have also been utilized on occasion such as SUTRA. There are a growing number of hydrologic models that are capable of simulating integrated surface and sub-surface flow, such as HydroGeoSphere and MIKE SHE, but these have rarely been used to simulate saline floodplains.

The majority of DEWNR's groundwater models of the SA River Murray were developed to meet the requirements of the Basin Salinity Management Strategy, and therefore focus on the long-term impacts of land use change and salt interception schemes on River Murray salinity. With the exception of the Chowilla model, they are not designed to simulate floodplain processes in any detail.

A literature review identified seven existing groundwater models of saline floodplains which include potentially useful modelling approaches. These are the Chowilla floodplain model (Yan *et al.*, 2004; RPS Aquaterra, 2012), the Murtho SIS climate sequence model (AWE, 2010), the Lindsay-Walpolla (EM4) model (Aquaterra, 2009), the Shashe River Valley model (Bauer *et al.*, 2006), the Clarks floodplain MODFLOW model (Doble *et al.*, 2005), the Clarks floodplain HydroGeoSphere model (Alaghmand, 2014) and the cross sectional Chowilla floodplain inundation model (Jolly *et al.*, 1998).

The models differ in how they simulate floodplain processes. Evaporation, transpiration and recharge may be modelled as a combined function or as separate functions. A variety of different assumptions have been used to describe the evapotranspiration rate as a function of depth to water. Zones for ET and recharge may be defined based on vegetation, soil type, groundwater salinity, surface water model calculations of inundation area, and/or AEM data. Rates may be sourced from soil properties or unsaturated zone simulations. River level change may be averaged over different time periods and simulated using different MODFLOW boundary conditions. The models rarely simulate unsaturated zone processes directly or the impact of density differences due to salinity gradients.

Groundwater models can be informed by a large number of different types of data. Monitoring programs to support conceptual understanding and numerical groundwater models should depend on the aims of the project, budget and time.

Review of current approaches to surface water modelling

Surface water models include one and two-dimensional hydrodynamic models and hydrologic models. Many hydrodynamic river models rely on the one-dimensional solution to the Saint-Venant equation, or one of its simpler approximations for flow in the river. To model the floodplains, a quasi two-dimensional approach can be applied, in which the flooded areas are modelled as separate 1D river branches. Other models use a 1D hydrodynamic model for the main river channel, and a reservoir approach for the river floodplain. More complex models simulate river flows across floodplains using full two-dimensional solution to the river hydrodynamic equations. In contrast, hydrologic models use hydrologic routing approach to calculate flow, and rating curves are used to determine the relationship between river depth and flow rate. BIGMOD, IQQM, REALM and Source are also examples of hydrologic models. These types of models typically have short run times, and so are suitable for simulating large systems.

Hydrologic river models are frequently used to examine how changes in river management will impact on flows and salt loads. However, these river models do not include many important groundwater processes, or they include them only in highly simplified manner.

Trials of groundwater model approaches

MODFLOW with MT3DMS was selected as the groundwater modelling platform for reasons both conceptual and pragmatic. MODFLOW is capable of simulating the key floodplain processes and is also currently DEWNR’s preferred groundwater modelling platform. Given the short project timeframe and immediate need for groundwater modelling recommendations, it was preferable to build on DEWNR’s existing capabilities rather than introduce a new groundwater modelling platform at this stage.

A simplified MODFLOW/MT3DMS model of the SA River Murray floodplain was used to test different approaches to simulating three key processes: river level change (i.e. river-groundwater interaction), evapotranspiration, and recharge from floodplain inundation. Numerical experiments were performed to explore how process representation affects model outputs. Both groundwater flow and solute transport were simulated. The model is designed to represent generic conditions for the SA River Murray floodplain in the study area; where variations exist, the model uses parameters representative of Pike floodplain. Six base case models were developed, to cover a range of hydrogeological conditions. The river may have losing, throughflow or gaining conditions at low flow; there may be a river lock (i.e. change in weirpool height) in the centre of the floodplain, or there may be no lock. MODFLOW version, temporal discretisation, parameter sensitivity and process representation were explored.

MODFLOW2000 had numerical difficulties in simulating groundwater flow in the clay Coonambidgal Formation. MODFLOW2005 with the NWT solver and UPW package has a revised “rewetting” algorithm and was able to simulate groundwater flow in the Coonambidgal.

The numerical experiments show that using yearly averages of river stage and/or ET does not capture seasonal and climactic variations in potentiometric head. Monthly and sub-monthly stress periods do capture this behaviour. However, flux between river and groundwater is very sensitive to stress period length, particularly cumulative flux, so sub-monthly stress periods are preferable when flux is a key output.

Model results were sensitive to small changes in the representation of ET. ET parameters controlled whether a river was under gaining or losing conditions in low flow periods. Changes in extinction depth and rate altered the amplitude of hydrographs, while spatially-varying ET zones also altered the phase/timing of peaks and troughs in potentiometric head. A curved function of ET with respect to depth to water had lower overall actual ET rates than a linear function with the same extinction depth and maximum ET rate. Topographic detail also strongly influenced ET.

Freshening of groundwater by river water lowers groundwater salinity near the river, so a constant-salinity calculation may overestimate the salt flux in a dynamic river system.

Trials of surface water model approaches

This project has developed some simple routines for calculating salt loads to the river resulting from infiltration beneath inundated floodplain areas. These routines have been developed for possible inclusion in river hydrologic models that are used for water management purposes, and have been coded into a trial version of the Source river model. Although further testing of these routines is required before they are used for river management, initial testing results are been very positive. The inclusion of floodplain salt transport routines in a regional river flow model would allow the interaction between river flow, and flow regulation and salt loads to be examined.

Discussion and recommendations

MODFLOW and Source are two of a large array of possible choices of models. The choice of model will be influenced by the model aim, the key processes, and the accuracy with which the different models describe these processes. It will also be influenced by the time required to construct and run the model, and the ability to carry out the required number of simulations in the required time period. It will also be influenced by the familiarity of the modeller (and the modeller's institution) with different numerical codes.

Current approaches to modelling the lower River Murray and SA floodplains often involve the simulation of a single model simulating a single hydrological domain, without comparison to models of other domains. For example, the SA Salinity Register models of groundwater and the DEWNR/MDBA Source models of surface water are developed independently and their results are not compared to each other. For a region with considerable interaction between groundwater and surface water, one may also consider a multi-model approach, where separate models of different hydrological domains are co-developed to be consistent with each other. Discrepancies would identify gaps in conceptual understanding and in data. Another option is to use a fully-integrated model which simulates multiple domains simultaneously, which would require further data and the development of expertise in these model platforms.

Recommendations include lists of:

- Monitoring tasks which could inform modelling; a subset should be selected based on the model area and aims.
- Preliminary recommendations for groundwater modelling of saline floodplains; these are being trialled in Pike Floodplain model, which is currently under development by DEWNR.
- Further scenarios that could be run using the generic floodplain groundwater model.
- Further work on groundwater modelling that would require the use of other models and/or model platforms.
- Recommendations for further work on Source simulations.

1 Introduction

Juliette Woods, Linda Vears & Matt Gibbs

The River Murray is the principal river of the western Murray Basin in southeastern Australia. It is highly unusual as it is a major river system that flows through an extensive landscape with highly saline groundwater. The river is naturally prone to salinity, and this propensity has increased over the past century due to the construction of river locks and the introduction of large-scale land clearance and irrigation. River flow volumes and river level variability have been reduced, while the watertable has risen. The overall impact has been to degrade riverine ecosystems and increase river and floodplain salinity.

In South Australia (SA), the River Murray provides water for the city of Adelaide, numerous smaller towns, industry, stock, irrigation, and floodplain ecosystems; hence the management of river salinity is vital for the economy and the environment. In recent decades, State and Federal governments have invested in research, engineering works and evidence-based policy to control the salinity. This has been extremely successful, and the salinity of River Murray water in SA now meets legislative targets even under extreme conditions of drought and flood. Given this success, the focus of salinity management has widened to include the management of floodplain salinity, to improve the health of riparian ecosystems.

Management of the lower River Murray floodplain requires a detailed understanding of its hydrological and hydrogeological processes, including those which mobilise or store salt. Much research has been undertaken to monitor, conceptualise and simulate the dynamics of floodplain salinity. However, there is currently no consistent and comprehensive approach to modelling the interaction of surface water and groundwater in the lower River Murray floodplains.

To address this need, in 2014 the Goyder Institute for Water Research commissioned the study *Modelling salt dynamics on the River Murray floodplain in South Australia*, a collaborative research project with contributions from Flinders University, CSIRO Land and Water, and the Department of Environment, Water and Natural Resources (DEWNR). The project consisted of two main tasks. Task 1 was a review of data and literature on salinity risk assessments and floodplain modelling in the lower River Murray. Task 2 developed and tested methods of simulating the salinity dynamics of the lower River Murray floodplain.

The companion report, *Modelling salt dynamics on the River Murray floodplain in South Australia: Conceptual model, data review and salinity risk approaches* (Woods, 2015a), documents the resulting conceptual model, data review, and salinity risk methodology discussion. This report documents the literature review and testing of modelling approaches. A third report contains the Appendices (Woods, 2015b).

1.1 Study area and scientific context

The study area consists of the floodplains of the River Murray in South Australia, from the Border to Morgan. This is within the South Australian part of the Murray Basin (Figure 1-2). Downstream of Morgan, there is little risk of salinity because groundwater salinities are lower and there is little flow from groundwater to the river. The Coorong, Lower Lakes and Murray Mouth (CLLMM) region is not included as its hydrology and hydrogeology are significantly different. While the study area is within South Australia only, similar salinity dynamics exist as far upstream as Nangiloc-Colignan in New South Wales and Victoria. More broadly, similar dynamics may be seen for any freshwater river that interacts with saline groundwater. Hence the literature review includes studies of sites in SA, interstate and overseas.

The SA River Murray lies within the westernmost part of the Murray Basin. The geology and hydrogeology is summarised in Woods (2015a).

The dry climate is responsible for the high salinity of the regional groundwater, concentrating the small proportion of salt in rainfall over tens of thousands of years (Herczeg *et al.,* 2001). The regional groundwater is typically 20,000 mg/l but can be above 100,000 mg/l (Telfer *et al.,* 2012).

The lower River Murray is a linear oasis within this dry landscape. The river has carved a trench through the upper sediments of the Murray Basin, dividing the topography into floodplain and “highland”. Within the trench, some freshwater is provided by rain, but the majority is delivered by the River Murray via surface channels, episodic flooding, and by freshening groundwater. The freshwater supports a diverse ecosystem, most strikingly the riparian river red gums and black box eucalypts.

The movement of freshwater and salt within the floodplain is complex. The river brings freshwater from upstream. Regional groundwater flows into the floodplain sediments, bringing salt. Evapotranspiration concentrates salts in the soils and groundwater. Flow between the river and floodplain aquifer depends on the gradient between the river level and the watertable, which changes over time and may reverse due to complicated interacting factors. Anabranches and wetlands may store and release salt. The system is extremely dynamic, and responds strongly to climatic conditions such as drought and flood. Woods (2015a) summarises what is known about floodplain hydrological processes which mobilise or store salt.

The system is further complicated by anthropogenic change. The construction of river locks in the 1920s/1930s altered the balance between river levels and groundwater levels, and changed some anabranches and wetlands from ephemeral to permanent while drying out others. Large-scale irrigation withdraws substantial volumes from the river. Land clearance and irrigation have increased recharge to groundwater, raising the watertable and mobilising more regional salt to the floodplain. Overall, the impact has been to reduce the volumes of freshwater flowing into the SA River Murray floodplains while increasing the volumes of saline groundwater flowing in. The average and peak salinity of the lower River Murray increased and floodplain vegetation was damaged.

Numerous works to control salinity have been undertaken. Some aim to reduce the flow of regional saline groundwater into the floodplain trench. These include controls on land clearance, the rehabilitation of irrigation areas to minimise recharge to the watertable, and the construction of Salt Interception Schemes (SIS). Other works alter the flow of freshwater within the floodplain, using regulators, weir pool manipulation, wetland management, artificial/environmental watering and pumping from the floodplain aquifer.

Works to control regional long-term salinity have been informed by numerical models, which estimate the impacts of management options. Regional-scale models of surface water include MSM-BIGMOD and Source. Regional groundwater models include the SIMRAT rapid assessment tools and the more detailed Salinity Register MODFLOW models. These models have been developed by or for DEWNR and the Murray Darling Basin Authority, and are well-tested and comprehensively reviewed. They employ consistent assumptions across the region.

Floodplain salinity control works have also been modelled, albeit on a case-by-case basis. However, in many situations the salinity impacts of possible management actions are difficult to quantify due to limited data, the complexity of the dynamics, and a lack of a standard approach. In these situations, the salinity impacts may instead be evaluated using a risk assessment method.

There is a need to improve the simulation of salt dynamics in the SA River Murray floodplain. This will improve our understanding and management of the floodplain and river. There are numerous prior studies to learn from which can be trialled and evaluated. From this a consistent preliminary approach can be derived and later improved upon incrementally.

Woods (2015a) concluded that there are several inter-linked challenges in modelling floodplain salt dynamics. The dynamics are controlled by numerous processes and drivers, many of which interact, and some of which are poorly understood. The processes and features may not be simulated well by standard codes. The dynamics impact domains which are usually modelled separately: groundwater, unsaturated zone, surface water, and atmosphere. Each domain requires specialist expertise and employs assumptions and conceptualisations that may differ from those used in other domains. Each process will vary spatially, depending on conditions and on heterogeneity. Each process will vary over time, some on a daily basis, others seasonally or over much longer timeframes (Telfer *et al.,* 2012). Finally, there may be insufficient data to characterise the floodplain of interest.

Nevertheless, scientific advances continue to be made and policy changes (Section 1.2) require the improvement of modelling methods. Quantifying the salinity impacts on the river and floodplain is the primary challenge for the next generation of SA River Murray groundwater models.

1.2 Policy context

Schedule B to the Murray-Darling Agreement establishes a requirement to identify, assess, report, monitor and review management actions which cause a significant long term increase in the salinity of the River Murray at Morgan. These actions must be entered onto the Murray-Darling Basin Salinity Registers.

The Basin Plan expands obligations to manage and report on short term salinity and water quality targets for the River Murray. All River Operators and Environmental Water Managers must have regard for these targets when making flow management decisions. Having regard for water quality targets must be done in the context of the outcomes of the Basin Plan, to deliver environmental flows to help protect and restore River Murray wetlands and floodplains while maintaining water quality for all water users.

To fulfil these requirements and to provide assistance to policy makers and environmental managers, South Australia is seeking to improve understanding of the short-term movement of water and salt within the floodplain landscape, under present flow conditions and under various alternate options for delivering environmental water.

1.3 Current SA government models of River Murray hydrology and hydrogeology

1.3.1 Groundwater models

The majority of DEWNR's groundwater models of the SA River Murray were developed to meet the requirements of the Basin Salinity Management Strategy (BSMS), which is discussed in Section 1.2. The BSMS focuses on the long-term impacts of land use change and salt interception schemes on River Murray salinity. Estimates of these salinity impacts are recorded and reported in the BSMS Salinity Registers. Since the BSMS was agreed, South Australia has developed a series of groundwater models to estimate salinity debits and credits for the Registers. The models calculate the groundwater flux and salt loads to the floodplain edge and/or the river. The scale and design of these models is suitable for estimating long-term, regional impacts.

The models are of two types: the rapid assessment tool SIMRAT and a suite of MODFLOW models.

SIMRAT was developed for the Murray-Darling Basin Commission (MDBC) and accredited in 2005 to assess the salinity impacts of new irrigation (Fuller *et al.,* 2005). It is programmed within a GIS framework, employing analytical equations and geographically-distributed parameters. It calculates the change in flux and salt load from the regional aquifers to the edge of the floodplain with the assumption that salt entering the floodplain enters the river. This consistent and deliberately simple approach can be used in areas where there is a high uncertainty in the hydrogeological factors which influence groundwater salt flux to the river. It is used in New South Wales and Victoria as well as South Australia (Woods *et al.,* 2015).

The MODFLOW models are collectively known as the SA Salinity Register models. They cover the following reaches of the SA River Murray:

- Chowilla floodplain, including areas in New South Wales, South Australia and Victoria (Yan *et al.,* 2005)
- SA Border to Lock 3, which includes the following sub-models:
    - Pike-Murtho (Woods *et al.,* 2014)

- Berri-Renmark (Yan *et al.,* 2007)
- Loxton-Bookpurnong (Yan *et al.,* 2011)
- Pyap to Kingston (Yan *et al.,* 2008)

- Woolpunda (Woods *et al.,* 2013)
- Waikerie to Morgan (Yan *et al.,* 2012)
- Morgan to Wellington (Yan *et al.,* 2010).

These models have been used to assess impacts of native vegetation clearance, irrigation, improvements in irrigation practice and infrastructure and the SIS. The Chowilla model has been used to assess the impact of a new regulator on Chowilla Creek, so is designed somewhat differently. The Salinity Register models calculate the groundwater flux and salt load to the SA River Murray. Each model is based on detailed hydrogeological information. They are calibrated against observed potentiometric heads and other outputs are compared against supplementary data sources, such as geophysics and in-river salinity surveys. Yearly timesteps are used. The grid size is typically 125 m. River levels and evapotranspiration (ET) generally do not change over time; the exception is the Chowilla model, which is described in greater detail later in Chapter 2. The models simulate groundwater flow but not solute transport. DEWNR groundwater models of the SA River Murray are not connected in any way with surface water models of the SA River Murray (the Chowilla model is again a partial exception).

The models and reports must be reviewed both internally and by external reviewers; if approved, they are said to be "accredited" by the MDBA. In accordance with Schedule B of the Murray-Darling Basin Agreement, all models which support salinity impact assessments are required to undergo a review at intervals of no more than 7 years. The SA Salinity Register models have generally been revised, updated, recalibrated and re-accredited every five years. As such, they embody the latest hydrogeological information. The SIMRAT model has not been revised since its inception, although a recent review makes recommendations to improve and update it (Woods *et al.,* 2015).

DEWNR also develops other MODFLOW models for site investigations. An example is the cross-sectional model of Clark's Floodplain in Bookpurnong (Berens *et al.,* 2009).

DEWNR is a government body which needs to make defensible decisions based on reliable software. At present, DEWNR's groundwater modelling expertise is concentrated in MODFLOW, but includes MT3DMS, Excel spreadsheets, GIS, Python and R. MODFLOW remains the primary platform as it is the industry standard: robust, well-tested, well-supported by its developers and third-party GUI developers, and with an open source code. Other, more specialist codes generally simulate more complicated physics, solving nonlinear equations which are prone to numerical instability. As such, DEWNR will continue to use MODFLOW as its first choice in groundwater model platforms, but may employ other models where the physics requires it.

1.3.2 Surface water models

The Science, Monitoring and Knowledge Branch within DEWNR currently use a wide range of surface water models of the SA River Murray and environs. Coupled 1D and 2D models are used to represent the floodplain anabranches of Chowilla Creek, Pike River and Katarapko Creek (e.g. McCullough, 2013). Full 2D hydrodynamic approaches are used to represent inundation in each weir pool, for example for assessing weir pool raising scenarios (Macky & Bloss, 2012). Hydrologic models, such as the BIGMOD and Source models of the SA Murray developed by the MDBA, are regularly used to assess the influence of barrage operations on water levels and barrage flows based on a flow forecast at the SA border and a range of diversion and loss scenarios.

However, these models either do not simulate salinity impacts at all, or do so in a limited sense. It is possible to model the transport of salt (advection and dispersion) in the hydrodynamic models, however this has not been undertaken to date. This approach also requires fixed salt inputs to be applied, surface water – groundwater interactions are not modelled explicitly currently. For the hydrologic models, salt transport is modelled in the same way as BIGMOD, an approach which adopts fixed salt inputs and does not provide any predictive capability to represent the change in river operations or flow delivery scenarios on in river salinity. More and more, river operations are being undertaken along the river to enhance environmental outcomes, for example Chowilla regulator operation and weir pool raisings at Lock 1 and Lock 2. The ability to simulate the potential effect of these types of operations on river salinity within the models currently used is highly desirable to inform management decisions.

1.3.3 Need for improved capabilities

The current generation of Salinity Register models are suitable for evaluating long-term, regional-scale impacts on groundwater salt loads to the SA River Murray. They simulate regional hydrogeological processes, a low-flow River Murray, and include a simple representation of the floodplain. MDBA reviewers have praised the models, but note shortcomings of the overall conceptualisation. The omission of river and floodplain dynamics makes it easier to analyse and interpret the model results, which is desirable given the models' aim of Salinity Register accounting, but may lead to systematic biases in calibrated parameters. The model developers and reviewers have raised this repeatedly in recent years with the MDBA, but the consensus was to continue with the current conceptualisation until sufficient resources were available to develop and trial a next generation model that included river and floodplain dynamics.

Another reason for considering floodplain dynamics is that there have been changes in the management of lower River Murray floodplains. A variety of techniques are being trialled to improve the health and robustness of floodplain ecosystems. Weirpool manipulation, new regulators, artificial/environmental watering, and floodplain pumping are being introduced. Examples include: the trials conducted at Bookpurnong in 2006, the Chowilla regulator, Riverine Recovery Program works, and the seven-year SA Riverland Floodplains Integrated Infrastructure Program (SARFIIP) which focuses on Pike and Katarapko floodplains. Groundwater models of the floodplain would assist with the engineering design of such works and with the evaluation of their salinity risks.

Finally, the Murray Darling Basin Plan expands obligations to manage and report on short-term salinity and water quality targets for the River Murray. The current Salinity Register models are not designed to simulate short-term impacts.

For these reasons, South Australia is seeking to improve understanding of the short-term movement of water and salt within the floodplain landscape, under present flow conditions and under various alternate management options. It is likely that this will require the development of two types of models: rapid assessment tools and detailed site-specific models. The rapid assessment tools could quickly provide rough estimates of salinity impacts in areas where there is little data and/or little risk. The detailed, site-specific models could be used to aid engineering design and to evaluate salinity impacts where risks are higher or where greater detail is needed.

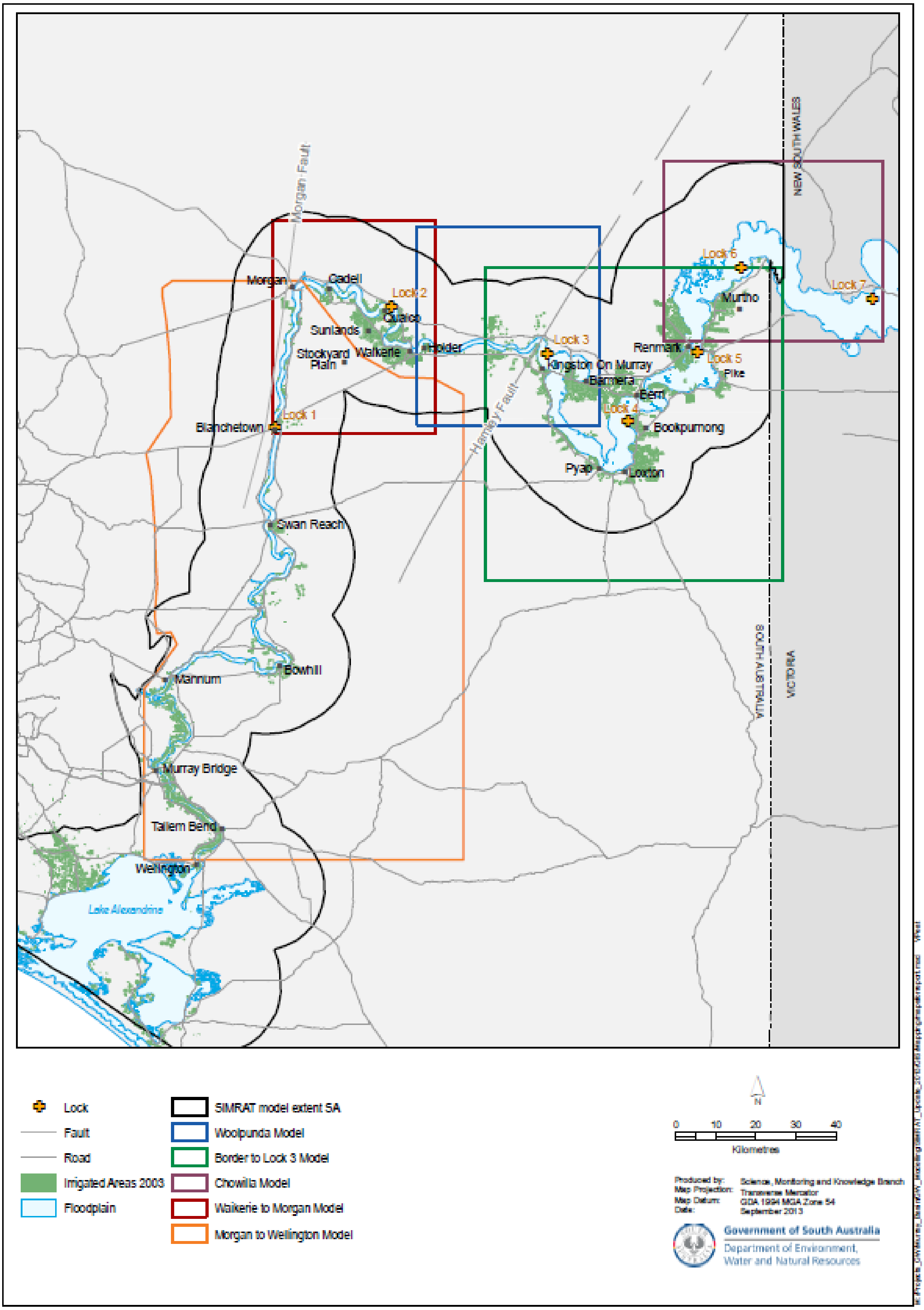


Figure 1-1 Extent and location of DEWNR's groundwater models for the SA MDB

1.4 Project aims

The Basin Plan increases State obligations to manage and report on salinity and water quality targets for the River Murray. A major outcome of the Basin Plan is to deliver environmental flows to help protect and restore River Murray wetlands and floodplains. To do so effectively, SA must understand the short-term movement of water and salt within the floodplain landscape, under present conditions and under various management options for delivering environmental water. Short-term salinity impacts, both positive and negative, should be calculated in a way consistent with the requirements of the Basin Plan and Schedule B. Current tools do not simulate the floodplain in the detail needed.

The project provides foundational knowledge, data and outcomes for existing and emerging environmental programs, including the Murray Futures Riverine Recovery Program, the South Australian Riverland Floodplain Integrated Infrastructure Program (SARFIIP) and future salinity management activities.

The companion report (Woods, 2015a) reviews the geology and hydrogeology of the Murray Basin, including details of the SA floodplains of the study area. It presents a conceptual model based on a literature review of floodplain salinity dynamics, identifying relevant physical processes, and the natural and anthropogenic drivers which impact the processes. Woods (2015a) also reviews the available datasets and approaches to salinity risk assessment.

The key outputs presented in this report are:

- Literature review of approaches to simulating the dynamics of saline floodplains. Chapter 2 discusses models focussed on groundwater dynamics, including a discussion of relevant datasets. Chapter 3 considers modelling approaches that are focussed on the surface water system.
- Key floodplain processes are examined to determine how these can be simulated (Chapter 4 and 5). This includes assessments of the impact on simulation accuracy of different assumptions/simplifications and data limitations.
- MODFLOW and Source pilot models are developed and tested to simulate lower River Murray and floodplain dynamics in Chapters 4 and 5 respectively.
- A works program is developed to prioritise improvements in the modelling required and inform targeted data collation and scientific studies (Chapter 6).

The Appendices provide the Python and Fortran programs developed as part of this project. A further Appendix provides results from the groundwater simulations. The modified Source program is not included in the appendices as the Source program is a proprietary code owned by eWater.

Due to the breadth of the modelling required, feedback was sought from modelling experts and floodplain management stakeholders. In February 2015, preliminary numerical models were presented to groundwater modellers at a workshop and were subsequently discussed at a Policy Advisory Committee meeting. Details of those consulted are provided in the Acknowledgements.

The key overall research outcome is how to represent floodplain processes to inform floodplain and river salinity management, including estimates of the uncertainty introduced by model assumptions. The outputs will enable the progressive development of models of the SA River Murray to simulate the impact of environmental actions on groundwater flow and salinity, including exchanges with the River Murray and freshening of floodplain aquifers. They will also inform models of other regions where surface water – groundwater interactions are important.

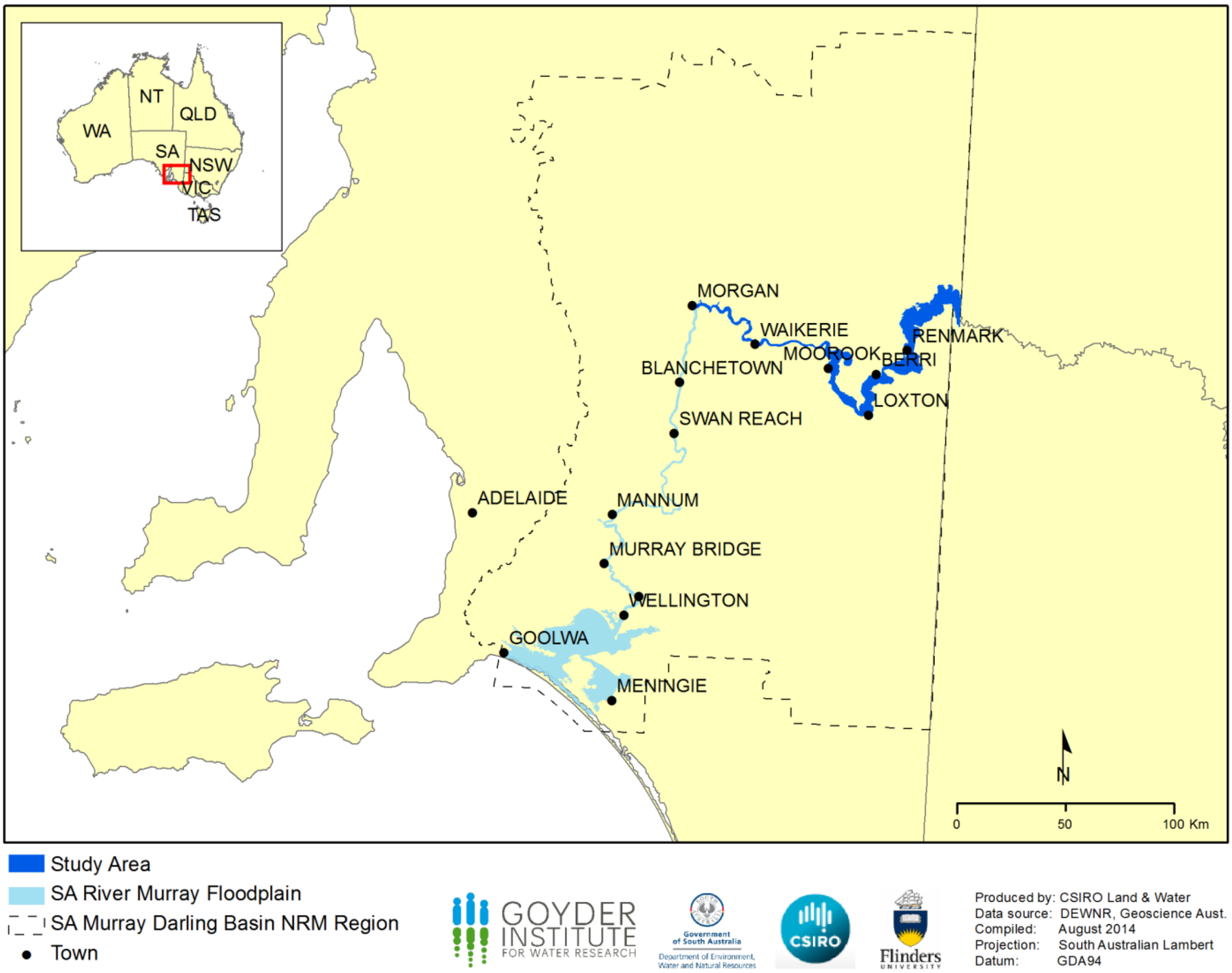


Figure 1-2 Location of the SA Murray Basin area

2 Groundwater modelling approaches

Tariq Laattoe, Juliette Woods, Chris Li & Virginia Riches

To simulate the water and salt dynamics in the River Murray floodplain, we need to understand the geological context, develop a conceptual model of the floodplain dynamics and review the available data. The companion report, Woods (2015a), provided this overview and identified key processes as:

1. Regional groundwater inflows
2. Evapotranspiration
3. River level change
4. Floodplain inundation, natural
5. Floodplain inundation, artificial
6. Evaporation from surface water

This chapter identifies methods used in the literature to represent these processes in numerical groundwater models. Section 2.1 discusses key floodplain groundwater models which include potentially useful modelling approaches. Section 2.2 considers model software. Section 2.3 considers each key process in turn, and compares different ways of simulating them. Section 2.4 considers which datasets are potentially useful at the different stages of numerical model development. Finally, Section 2.5 makes recommendations for testing and adopting simulation techniques. Some of the recommended techniques are tested in Chapter 4 while others relate to works that would need to be addressed in parallel and subsequent projects, as discussed in the report recommendations (Chapter 6).

2.1 Exemplar floodplain models

Exemplar models featured in this section have been selected on the basis of their simulated processes, those identified as significant factors for salt mobilisation on the floodplains of the Lower Murray. We do not present all model input parameters as our aim in this section is to highlight unique approaches used in the selected models. Further details of the models are located in their referenced publications. The selected models are:

1. The Chowilla floodplain model (Yan *et al.*, 2004; RPS Aquaterra, 2012)
2. The Murtho SIS climate sequence model (AWE, 2010)
3. The Lindsay-Walpolla (EM4) model (Aquaterra, 2009)
4. The Shashe River Valley model (Bauer *et al.*, 2006)
5. The Clarks floodplain MODFLOW model (Doble *et al.*, 2005)
6. The Clarks floodplain HydroGeoSphere model (Alaghmand, 2014)
7. Cross sectional Chowilla floodplain inundation model (Jolly *et al.*, 1998)

2.1.1 The Chowilla floodplain model

The Chowilla model is one of the SA Salinity Register models (see Section 1.3) and was developed using MODFLOW by Yan *et al.*, (2004). The floodplain at Chowilla is well-recognized as a discharge area for regional saline groundwater to the River Murray during floods. The development of the locks and weirs has increased the adverse salinity impacts to the floodplain and the water quality of the River, including the anabranch complex.

Initially, the Chowilla model was similar to the other SA Salinity Register models in that it was used to simulate the regional aquifer system under low flow River conditions and did not include the simulation of any flooding. The model has been updated multiple times since its initial development and has been revised to assess the benefits and impacts of various management schemes proposed to improve floodplain and River health. It is now considered the benchmark floodplain model in SA. The data density available in the Chowilla region sets this model apart from any of the other floodplain models considered in this section.

In 2005 to 2006 the model was used for the first time to simulate aquifer hydraulic response to flooding, both natural floods and floods induced by the proposed Chowilla Regulator (Figure 2-1). The modelled results were used to estimate groundwater salt load delivery to River and were also used by CSIRO to predict vegetation health response (Overton, 2005). In 2007, the model was used to simulate aquifer response to flooding from a 30 year River Murray flow hydrograph (Howe *et al.*, 2007). Fundamental model parameters were not altered from those used in Yan *et al.* (2004) other than to simulate flooding.

The 2012 update (RPS Aquaterra, 2012) comprised several changes to model input parameters including:

1. Updated LiDAR topography which has resulted in changes to surface topography, flood inundation area, stream levels and evapotranspiration.
2. Refinements of the groundwater salinity zones based on AEM data.
3. Clarification and documentation of the flood inundation recharge rates.

The model covers an area of approximately 55 km (east-west) by 45 km (north-south) and includes the western part of Lake Victoria and the entire Chowilla floodplain. Minimum grid size is 76.5 m by 62.5 m on the floodplain with maximum grid size of 305 m by 250 m near the model edges. It comprises 5 layers, 2 of which are combined floodplain plus adjacent regional units and the remaining 3 layers form the regional aquifers beneath the floodplain. Initial development in 2004 featured constant head cells for River and anabranch representation. The 2012 update has replaced these with river cells (a type of head dependent flux hydraulic boundary cell) from MODFLOW's River package. Variable streambed conductance for the anabranch creeks has also been implemented to simulate the effects of increased wetted perimeter with increased water level. Evapotranspiration rates implemented in the model range from 150 to 250 mm/yr (much less than 1 mm/d) with extinction depths between 1.5 and 3 m. Potential ET for the area is given at 2,000mm/yr.

Simulating flood response in the Chowilla groundwater model requires estimates of (i) the extent of inundation for a range of specified flows, (ii) River Murray and creek levels during and after floods, and (iii) recharge rates for areas under inundation. The first two items were provided by a hydrodynamic model developed by Water Technology (2009). Recharge rates for inundated areas were mapped to spatial zones provided by CSIRO after Overton *et al.* (2005) which are based on soil profiles. The recharge zones ultimately input into the model are a combination of the inundated area for a specific River flow and the inundation recharge zones (Figure 2-2). Recharge rates were supplied by Overton *et al.*, (2005) but were significantly lowered during model calibration.

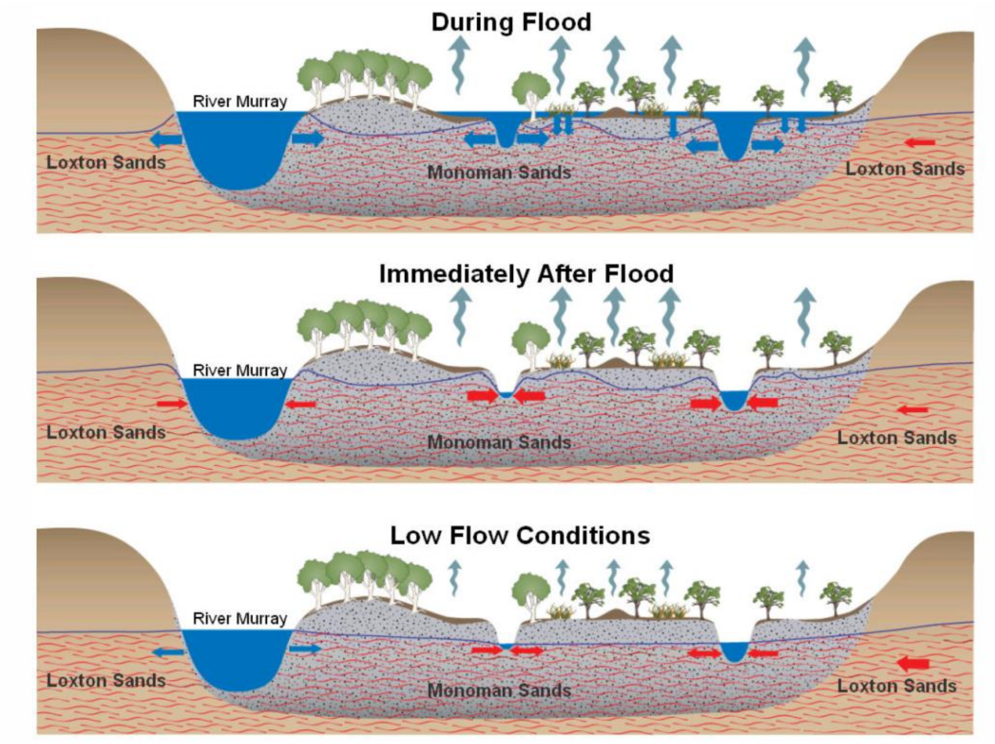


Figure 2-1 Conceptual model of the groundwater responses during and post flooding of the Chowilla floodplain

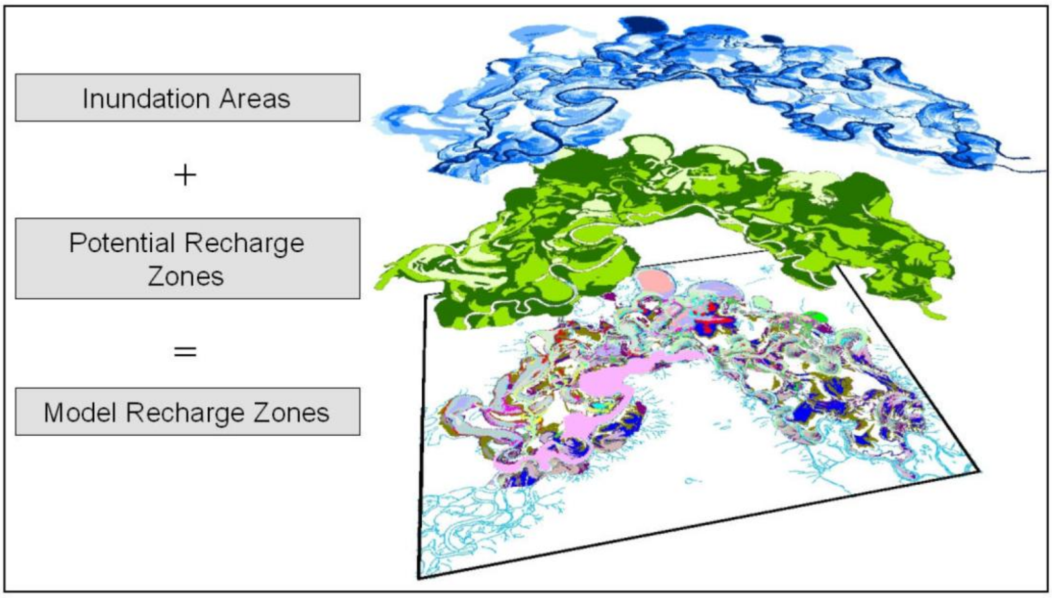


Figure 2-2 Information layers combined to define the model recharge zones (Source: Aquaterra)

The mean groundwater salinity in both the Chowilla 2007 and Chowilla 2012 models is similar (~25,000 mg/l), however Chowilla 2007 has a single value applied to the entire model whereas Chowilla 2012 has a spatially variable salinity based on AEM data (~25,000 mg/l on the western half of the Chowilla floodplain and ~35,000 mg/l on the eastern half). Other notable assumptions and simplifications of the model include:

1. A flat River Murray weir pool (no backwater curves).
2. A simplified River Murray flow hydrograph comprising monthly time steps and 20,000 ML/day flow magnitude divisions with a river flow of 5,000 ML/day assumed between flood events.
3. Inundation areas and stream levels for flows of 80,000 and 100,000 ML/day are represented by pre-locking conditions in the hydrodynamic model.
4. Anabranch bed conductance parameter values that increase as the water level in the anabranches increases (representing increased wetted perimeter).
5. Evapotranspiration from deep-rooted floodplain vegetation in the flush zone represented by an area of negative recharge.

Notwithstanding the conceptual uncertainties and knowledge gaps outlined above, the good match to measured groundwater levels and salt loads achieved by the model would indicate that the level of model simplicity that has been adopted is adequate. Moreover, the simplifying assumptions are all considered to be conservative as they are inclined to result in over-prediction of salt load to River. However, the time steps dictated by the simplified hydrograph used for river stage means that short term (sub monthly) events cannot be simulated.

2.1.2 The Murtho climate sequence model

As part of the design of the Murtho SIS, AWE (2010) considered the impact of floodplain processes on SIS efficiency. A MODFLOW/MT3DMS numerical groundwater flow and solute transport model was constructed which simulated variable river levels and floodplain inundation. The model illustrates processes of flood recharge and recession, the impact of drought, and the mixing of fresh and saline groundwaters in the floodplain. ET was simulated more simply, in that potential ET rates were held constant over time.

The level of the River Murray was simulated over the BSMS Benchmark Period, i.e. from 1975 to 2010, including floods and droughts. Stress periods were hand-selected to match observed changes in river level, so that periods of rapid change had shorter stress periods. Stress periods ranged from 11 to 750 days in length. The river level was calculated for each stress period and each river kilometer based on observations from upstream of each lock and digitized backwater curves. The river was simulated using reservoir cells. When active, reservoir cells are identical to river cells except that they interpolate linearly between the level at the start of a stress period to the end of the stress period, so timesteps within the stress period more accurately reflect changing river levels. The base of the river was estimated from NanoTEM survey bathymetry data. Riverbed hydraulic conductivity was determined during calibration.

Overbank flow was also represented using reservoir cells. Each floodplain cell was assigned an elevation, based on LiDAR data, and a time series of river levels. The river levels assigned to a floodplain cell were those of the closest river kilometer (a simplification of flood hydrodynamics). A floodplain cell became wet (flooded) whenever the river level was higher than the elevation. The hydraulic conductivity of the floodplain reservoir cells was 10^{-3} m/day, representing the resistance of the Coonambidgal Formation.

Solute transport simulation assumed salinities based on the medians of observations. River Murray and flood waters were fresh (368 mg/l), irrigation recharge 500 mg/l, and the initial regional groundwater salinity was 28,930 mg/l. The groundwater salinity changed over time due to irrigation recharge, flows from the River Murray, overbank inundation, and ET.

The flow model was calibrated to hydrographs, the solute transport results were compared with geophysics data, and estimates of salt movement from groundwater to the Murray were compared with independent estimates from MDBA's BIGMOD model.

2.1.3 The Lindsay-Walpolla (EM4) floodplain model

[Note at the time of writing this section, no documentation on the EM4 model was made available by the Mallee Catchment Authority. The details included in this section are based on the contents of conference presentations by Hugh Middlemis at the GroundWater2010 conference.]

The Eastern Mallee 4 (EM4) model focuses on the Lindsay-Wallpolla floodplain downstream of the confluence between the Murray and Darling Rivers at Wentworth past Lock 10 to the SA border. The role of the model was to assist in making management decisions concerning environmental watering options under the MDBA's program 'The Living Murray". The sub-regional scale model was originally developed by Aquaterra in 2008 (Aquaterra, 2009) and further refined in 2010 to obtain detailed information on floodplain salt mobilisation processes. It uses MODFLOW-Surfact (a modified version of MODFLOW from Schlumberger) operating under Groundwater Vistas. It is designed as a medium to high complexity model, suitable for floodplain management purposes in terms of simulating the effects of environmental flows and related surface and groundwater interaction processes. The model results are used to inform vegetation health assessments, as well as for salinity management.

The Lindsay-Wallpolla floodplain is a data-rich area, which includes LiDAR, AEM depth slices, CSIRO soil recharge map [to be confirmed], vegetation health mapping, and 75 floodplain bores covering 15 transects perpendicular to the river (Figure 2-3). These data sources, when combined with a hydrodynamic model, allowed for detailed representation of floodplain processes.

The model has 100 x 50 m grid cells. Quarterly stress periods are used in which river stage and recharge are adjusted. River stage is assigned an average level based on pool level, backwater and anabranch levels. Two key processes that were required to be addressed to successfully model floodplain processes were floodplain inundation recharge and evapotranspiration.

Floodplain inundation was modelled as follows. Floodplain inundation areas were provided by the hydrodynamic model for a range of specified river flows. Floodplain inundation recharge rates were based on airborne electro-magnetic (AEM) data (Figure 2-4). AEM slices at 3 separate horizons were used to assign four salinity index classes. The product of the index values from each layer at any location is treated as a qualitative indicator of the rate of recharge. Recharge classes of 1, 3 and 6 mm/day were assigned (Figure 2-5), based on Overton *et al.* (2005) (pers com. H.Middlemis, 2014), similar to the Chowilla model, however the exact details are not currently available. The new recharge rates greatly improved calibration to potentiometric head levels observed during the more frequent periods of flooding during the mid-1990s.

The EM4 model also features spatially-distributed evapotranspiration rates and extinction depths. These were developed from the AEM data and also from vegetation health mapping developed by SKM. A greater maximum evapotranspiration rate was assigned to areas with denser and healthier vegetation, as obtained from analyzing National Vegetation Change data (NVC) and total dissolved salts (TDS). 14 classes were defined with maximum rates from 0 to 0.005 m/day implemented via the standard ET package in MODFLOW. This improved the model capability to match the recession periods for up to 5 years post-flooding.

The AEM slices were also able to provide information on interaction between the groundwater and the River Murray. The AEM datasets can identify losing and gaining conditions (similar to the NanoTEM) and determine the location and shape of freshwater lenses around the river. This provides another opportunity to check if the model replicated what was known about the system. The documents provided also identified that AEM was used to refine the calculation of salt load to the river in other models in the EM group (the EM2 suite), however no details of this process was included.

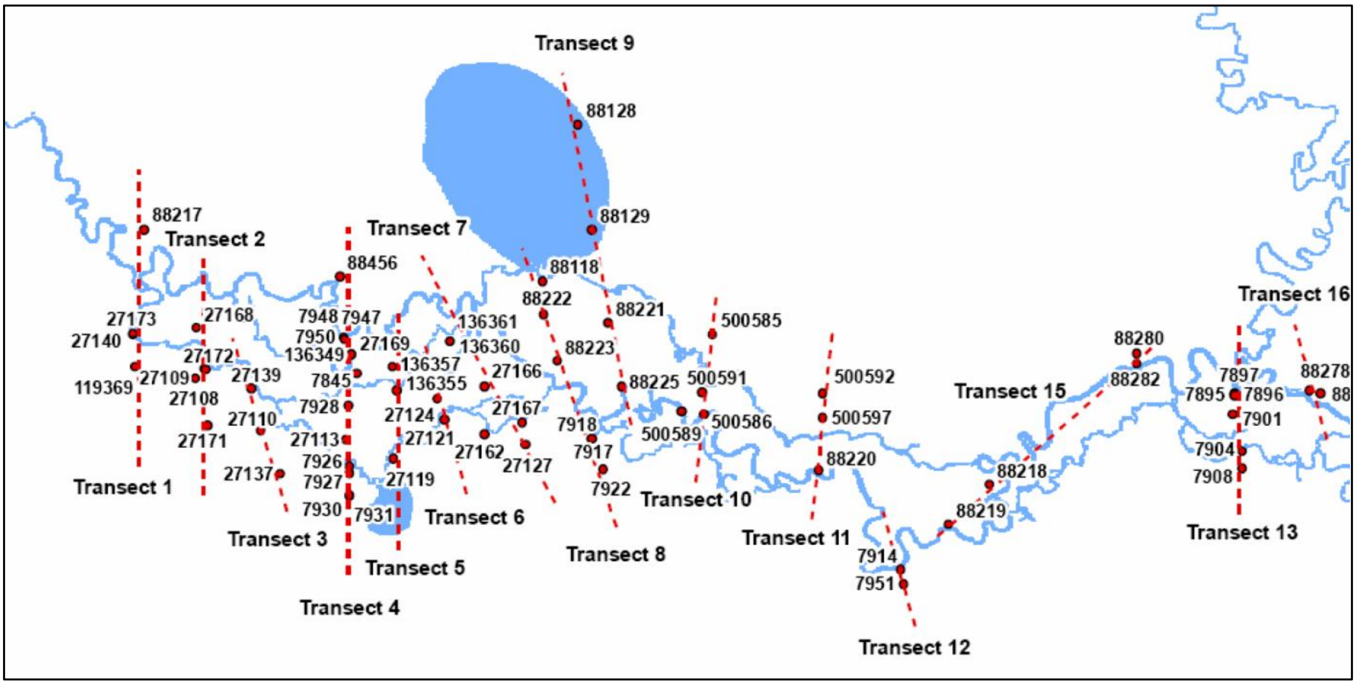


Figure 2-3 Bore transects used to provide head measurements for calibration of the EM4 model (Source: Aquaterra, 2009)

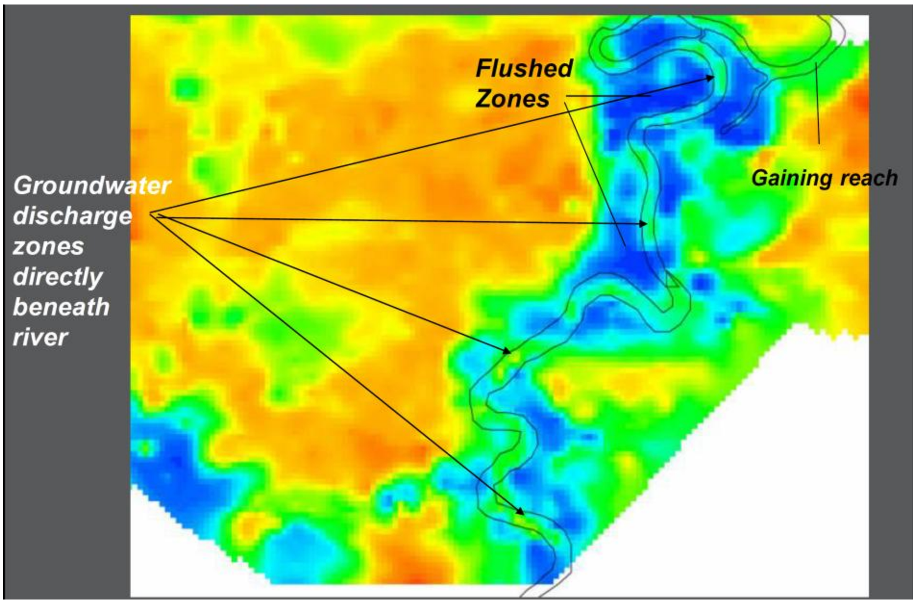


Figure 2-4 AEM slices identify flush zones and gaining reaches (Source: Aquaterra)

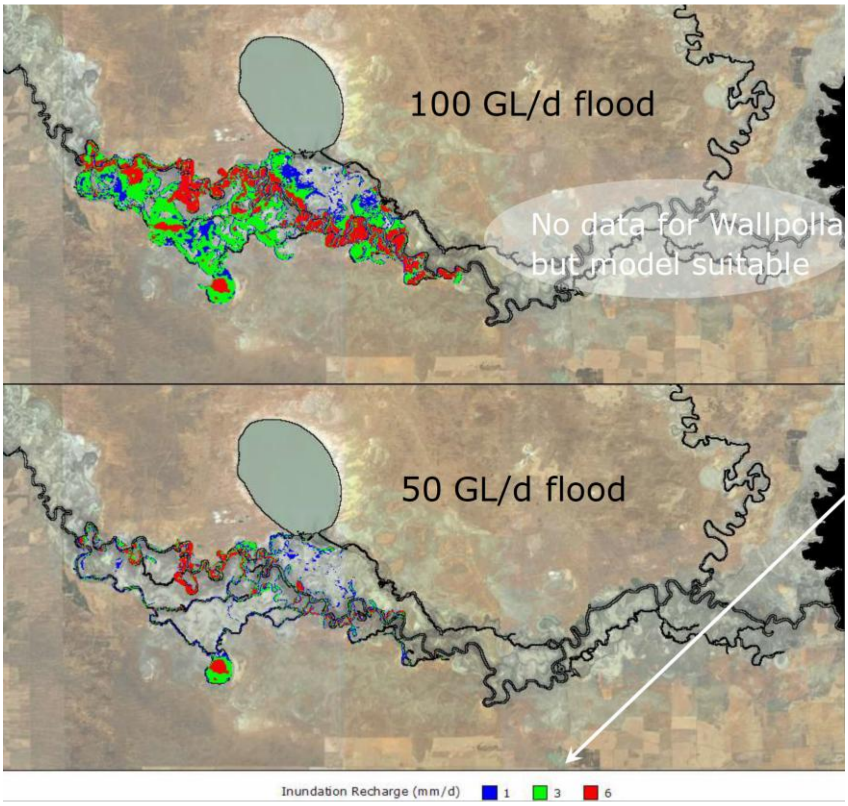


Figure 2-5 The spatially-distributed floodplain inundation recharge rates applied to the EM4 model (Source: Aquaterra)

2.1.4 The Shahse River Valley model

Bauer *et al.* (2006) present a model to simulate the formation of a large freshwater lens over centuries. It uses a modified evapotranspiration package in MODFLOW and SEAWAT. Similarities between the hydrologic settings in the lower River Murray and the Shashe River Valley make these modifications worthy of consideration for future modeling projects in SA.

The Shahse River Valley is located in northern Botswana at the south eastern fringe of the Okavango Delta. The regional climate is semi-arid with annual precipitation of 450 mm/yr. and potential evaporation of 1,500 mm/yr. The river valley is flooded every year when the Okavango Delta floods. One of the primary features of this river valley is a large freshwater lens that has been continuously exploited for domestic water supply. Reduced flooding and increased withdrawals has resulted in a dramatic decline in the water levels and mobilisation of the saline water (approximately 25 ppt.) from the deep saline aquifer surrounding the freshwater lens.

Similar to the floodplains of the lower River Murray, evaporation and transpiration in the Shashe River Valley are considered to be key processes that dominate the water budget. Ensuring that the behavior of these processes is accurately represented in a model is crucial.

Evaporation is mainly a function of depth to watertable, porous media properties and climatic conditions, while transpiration is a function of many variables including climate vegetation type, vegetation cover, depth to watertable and water quality. For a solute mass balance, the processes differ in the maximum concentrations produced. Evaporation can increase dissolved concentrations to an upper limit given the solubility of the salts while transpiration breaks down at significantly lower concentrations depending on the salt tolerance level of the plants.

In the standard SEAWAT software, evaporation and transpiration are lumped into a single process in the ET package described by 3 parameters only. R_{max}, the maximum rate of evapotranspiration, *S*, the water level elevation (given by the where the rate starts to decline which is usually the topographic surface, and *D*, the extinction depth where the rate is zero (Equation 2-1).

$$\begin{aligned}
&R = R_{\max} \quad \text{for } h \geq S \\
&R = R_{\max}\left(1-\frac{S-h}{D}\right) \quad \text{for } S-D \leq h \leq S \\
&R = 0 \quad \text{for } h \leq S-D
\end{aligned} \tag{2-1}$$

The simplest relationship for *R* between *S* and *D* used in most models is linear, as displayed above; however, recent investigations have shown that in most cases an exponential relationship better approximates the flux behavior. Through some relatively simple code modifications to the source code of the ET package in MODFLOW the following change can be implemented:

$$R = R_{\max}.\exp\left(-\frac{S-h}{D}\right) \quad \text{for } h < S \tag{2-2}$$

For the Shashe River valley simulation, this modification was not sufficient to reproduce system behavior and further modifications were made to separate evaporation from concentration-dependent transpiration. The linear relationship (Equation 2-1) was used for evaporation while the exponential (Equation 2-2) was used for transpiration. Transpiration was further modified to have a variable R_{max} value dependent on salt concentration given by the following:

$$R_{T,\max}(c) = \frac{R_{T,\max}(0)}{2}\operatorname{erfc}\left(\frac{c-cT}{\sqrt{2}\sigma}\right) \tag{2-3)}$$

Where $R_{T,max}$ (0) is the transpiration rate for freshwater (m/day), c_T is a threshold concentration(kg/m^3) and σ (kg/m^3) is a shape parameter describing the smoothness of the transition. Typical literature values for c_T range between 2 and 8 kg/m^3 (2,000 to 8,000 mg/l) and for σ between 0.5 and 4 kg/m^3 (500 to 4,000 mg/l). It should be noted that the form of the reduction function is hypothetical and is selected for mathematical convenience; notwithstanding, it is qualitatively similar to growth reduction functions commonly used in modeling literature.

Changes were also made to SEAWAT to reflect salt accumulation by evaporation and transpiration. For evaporation, the uptake of salt into the unsaturated zone will be related to the salinity of the groundwater, implying that salt will exit the modelled saturated zone. The significance of this effect is not made apparent in the results. Transpiration from the watertable will remove water but leave some or all the salt behind. In SEAWAT the respective source terms q_E and q_T are parameterised the following way.

$$q_i = \frac{R_i(c - c_{u,i})}{\Delta z} \quad \text{for } c \geq c_{u,i}$$
$$q_i = 0 \quad \text{for } c < c_{u,i} \tag{2-4)}$$

Where $c_{u,i}$ are the concentrations of the water taken up by evaporation and transpiration (kg/m^3) and Δz is the thickness of the cell from which the flux is drawn. In this manner, if the uptake concentration is equal to the aquifer concentration (c=$c_{u,i}$) then no accumulation occurs which is a reasonable approximation for the case of diffuse uptake of water linked to evaporation. If the uptake concentration is zero (c_u=0), all salt is left behind in the aquifer, which is a reasonable approximation for the case of diffuse uptake of water linked to transpiration.

These changes to the ET package can be made via modifications to the source code and were implemented because geochemical data used in the calibration process could not be matched without salt accumulation processes in specific areas. A similar condition exists on the floodplains of the Lower Murray and this approach is therefore worthy of further investigation when using models that include explicit representation of salinity. While physically more realistic than a lumped ET function, separation of evaporation and transpiration may only result in significant effects when simulation times range from decades to centuries.

2.1.5 The Clarks Floodplain MODFLOW model

This floodplain model determines which areas of the floodplain vegetation are most at risk from long-term salinity (Doble *et al.*, 2005). It is also able to show rates of lateral recharge from the River to vegetation. It provides a methodology for the spatial comparison of discharge.

Calibration was not included and precise input parameters from field data were not required. The methodology utilizes existing and tested evapotranspiration theory, and incorporates it into a combined recharge-discharge relationship that is implemented in MODFLOW. To develop a spatially-variable, depth-dependent function for recharge-discharge, the model requires spatial information for vegetation, soil and groundwater salinity, as well as flooding duration and frequency. The function is similar to the segmented evapotranspiration function of the MODFLOW ETS1 package but also incorporates a recharge component. Its use leads to more accurate long-term modeling of evapotranspiration and recharge processes. The referenced publication contains a significant amount of theory that was used to develop the combined recharge discharge relationship. We will present only the fundamentals and its implementation in MODFLOW in this report.

Clarks Floodplain at Bookpurnong was chosen as the study site, as increased irrigation in the adjacent highland has led to significant salt accumulation and vegetation dieback on the floodplain. The climate is semi-arid, with rainfall varying between 200 and 300 mm/yr, and potential evaporation of approximately 1,800 mm/yr. The floodplain is approximately 5 km wide and features ephemeral creeks and a lagoon at the break of slope to the highland.

Recharge on the floodplain is predominantly from flooding events where infiltration rates is presumed to approach a constant value. Recharge is therefore dependent on the vertical conductivity of the soil and the length of time that the surface is inundated. Equations are provided in Doble *et al.*, (2005) to calculate the recharge from a single flood event for certain soil parameters. Cumulative effects of multiple floods can therefore be made into a single function. The final recharge function is therefore able to account for recharge from multiple flood events without actually simulating the flood itself.

A method for calculating cumulative groundwater discharge through plant water uptake is also provided. Discharge in this method is governed by depth to watertable, soil texture and moisture content, and the ratio between groundwater chloride concentration and the threshold concentration for plant water uptake. The functions are combined through simple addition. They are represented conceptually in Figure 2-6. The function applies a net recharge or discharge to the model depending on whether the groundwater head is above the ground (maximum evapotranspiration), between the surface and equilibrium depth (point (a): net evapotranspiration), between the equilibrium depth and extinction depth (point (b): net recharge) or below the extinction depth (max recharge). The equilibrium point is where recharge and discharge fluxes are equal to zero in the long term. With this method it is possible to include varying recharge functions such as zero recharge when the groundwater is at the surface.

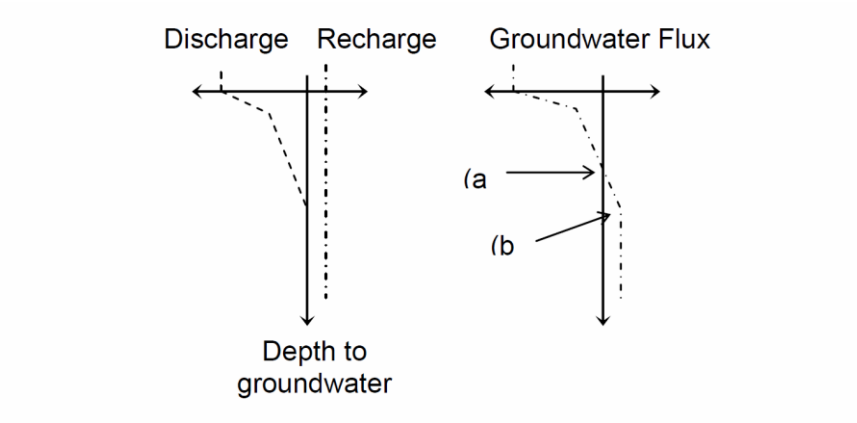


Figure 2-6 Conceptualisation of discharge and recharge in Doble *et al.* (2005)

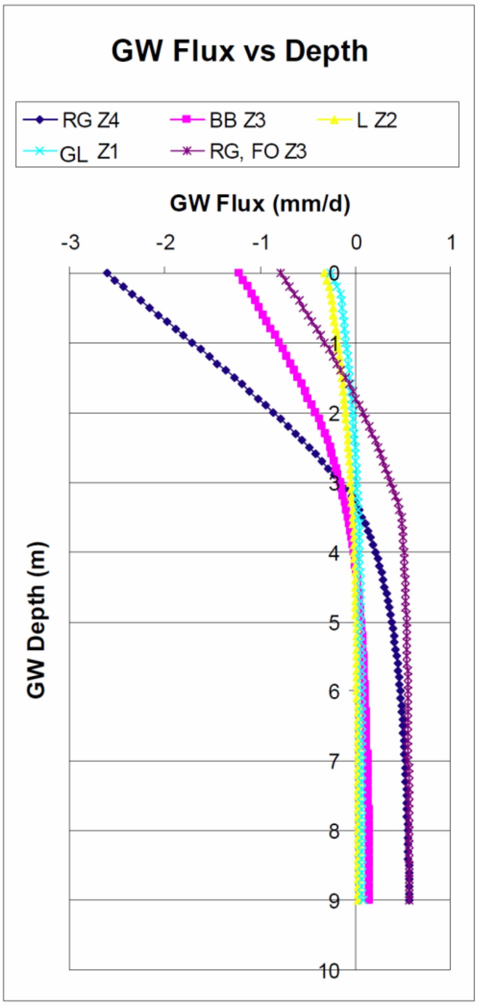


Figure 2-7 Groundwater flux vs depth for various scenarios (Source: Doble *et al.* 2005)

Figure 2-7 gives groundwater flux vs depth relationships for various floodplain vegetation soil and groundwater scenarios RGZ4 is Red Gums on clayey soil with low salinity groundwater. BBZ3 is Black Box on sandy clay with moderate salinity groundwater. L Z2 is Lignum on heavy clay with high salinity groundwater. GLZ1 is grassland on heavy clay with extremely shallow highly saline groundwater. RG FO Z3 is Red Gums with heavy clay flood out and moderate salinity groundwater.

The maximum ET and recharge rates are given in Figure 2-8.

There may be opportunities for this methodology in floodplain management scenarios featuring artificial inundation. The ability to simulate an actual flood or flooded area is a significant complication for model development. Spatial and temporal discretisation of the modelled domain requires careful consideration when attempting to capture the salt mobilisation processes from inundation scenarios. However, if the purpose of the model is to determine changes in flux to the River from regulator operations and environmental watering strategies over longer periods of time, then the combined recharge discharge approach provides a convenient method to incorporate inundation events into a net long-term spatially distributed flux and in doing so avoids the complications associated with explicit flood simulation.

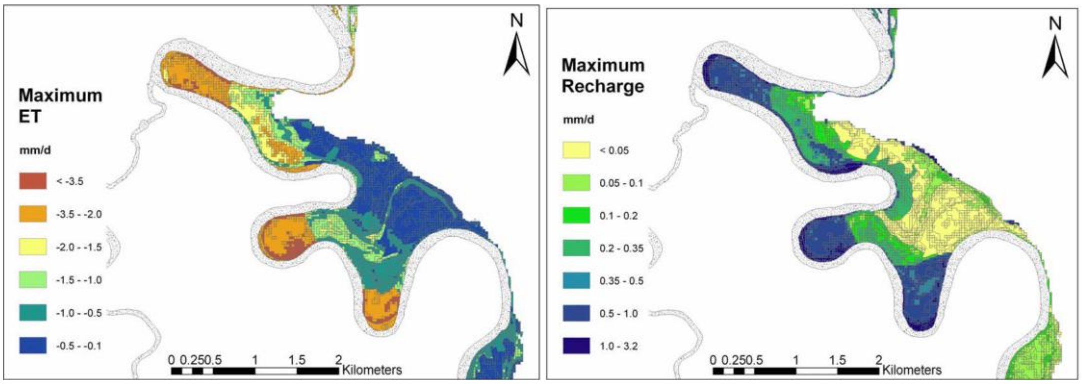


Figure 2-8 Maximum ET and recharge rates applied to Clark's Floodplain (Source: Alaghmand *et al.*, 2014)

2.1.6 The Clarks Floodplain HydroGeoSphere model

The main objective of this modeling investigation was to quantify the impacts of river stage manipulation on freshening of the shallow floodplain groundwater through bank storage (Alaghmand *et al.*, 2014). Motivation for the study was to inform policy of the true ecological impact of land management decisions specifically those involving weir pool manipulations. The study location was Clarks Floodplain, due to the density and abundance of data available for model input. HydroGeoSphere was identified as the model platform best suited to this study based on previous bank storage publications. A 3D transient flow and solute transport model was setup and run for a period of 2,070 days using a maximum time step of 1 day. The model spatial discretisation was based on a LiDAR Digital Elevation Model of the study site with a 10 m grid resolution and covered 61.3 ha of Clark's Floodplain from the floodplain break of slope to the River Murray main channel. Unlike the floodplain models developed with MODFLOW, this investigation included the unsaturated zone and explicit representation of the Coonambidgal Formation.

Figure 2-9 Boundary conditions for the HydroGeoSphere model of Clark's Floodplain. It also shows the modelled area and layering in a cross sectional projection along transect 1. Note yellow elements (Figure 2-9b) represent the Coonambidgal Formation, grey elements the Monoman Formation and purple elements the Upper Loxton Sands.

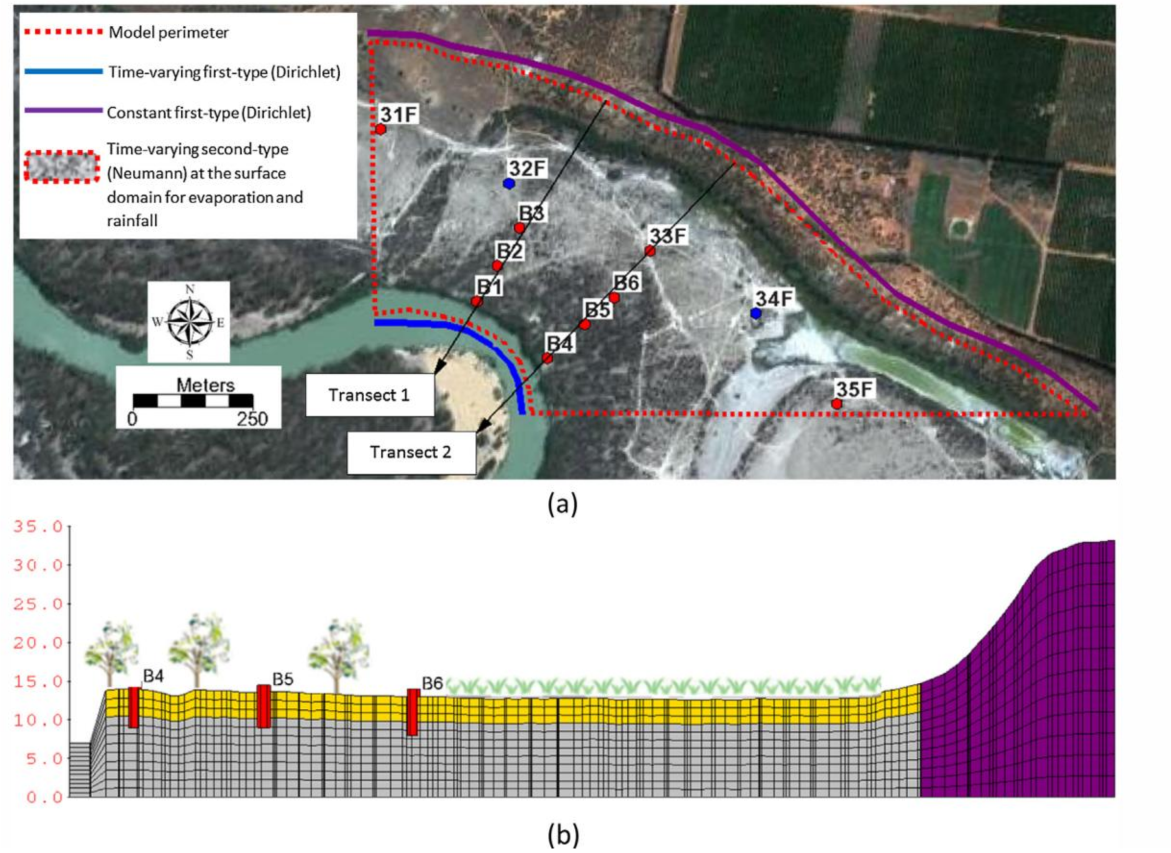


Figure 2-9 Boundary conditions for the HydroGeoSphere model of Clark's Floodplain (Source: Alaghmand *et al.*, 2014)

The model was manually calibrated to achieve an acceptable match to hydraulic heads observed at the wells and observed concentration patterns. The calibrated model was then used to run 12 river stage manipulation scenarios including variation in the size of the stage shift and the duration between manipulations. The model also simulated watertable drawdown by SIS bores currently located at the study site to determine the extraction rates necessary to prevent regional saline water from entering the floodplain.

The results of the study showed that raised river level stage at Clarks Floodplain will result in the development of a shallow freshwater band along the river extending approximately 100 m into the floodplain. The numerical experiments also showed that watertable lowering had the greatest impact on reducing the salinity of the floodplain. A follow-up investigation with the model includes floodplain inundation but the results of that study are in the publication process. One major limitation is that HydroGeoSphere does not simulate the evapoconcentration of salt in the groundwater or soils.

2.1.7 Cross sectional Chowilla floodplain inundation model

In this study by Jolly *et al.* (1998), the transport of salt between groundwater and streams was modelled and compared with available field data. The presence of large density contrasts between the fresher stream water and the saline groundwater results in density dependent flow behavior. The aim of this study was to use a variable-density flow and transport model to better understand the processes operating in the transport of saline groundwater to floodplain streams. The modeling platform SUTRA was selected due to its ability to model saturated-unsaturated conditions and variable-density flow and transport. Flood simulation also required time-varying boundary conditions which were relatively easy to implement on SUTRA in comparison to other modeling platforms at the time. The investigation did not attempt to match observed salt measurements exactly but was aimed at recreating observed system behavior.

The model simulates a 2D cross section as displayed in Figure 2-10. A transect of piezometers were installed at the site and sampled to provide data to constrain the model. The water quality of the piezometers did not change significantly after the flood except in the piezometers closest to the river and creeks, where salinity fell with the flood but rose quickly in the following months. A time-varying boundary condition is used to simulate the rise and fall of the water level in the River Murray in response to flood. Similar boundary conditions were used for Monoman Creek and Chowilla Creek.

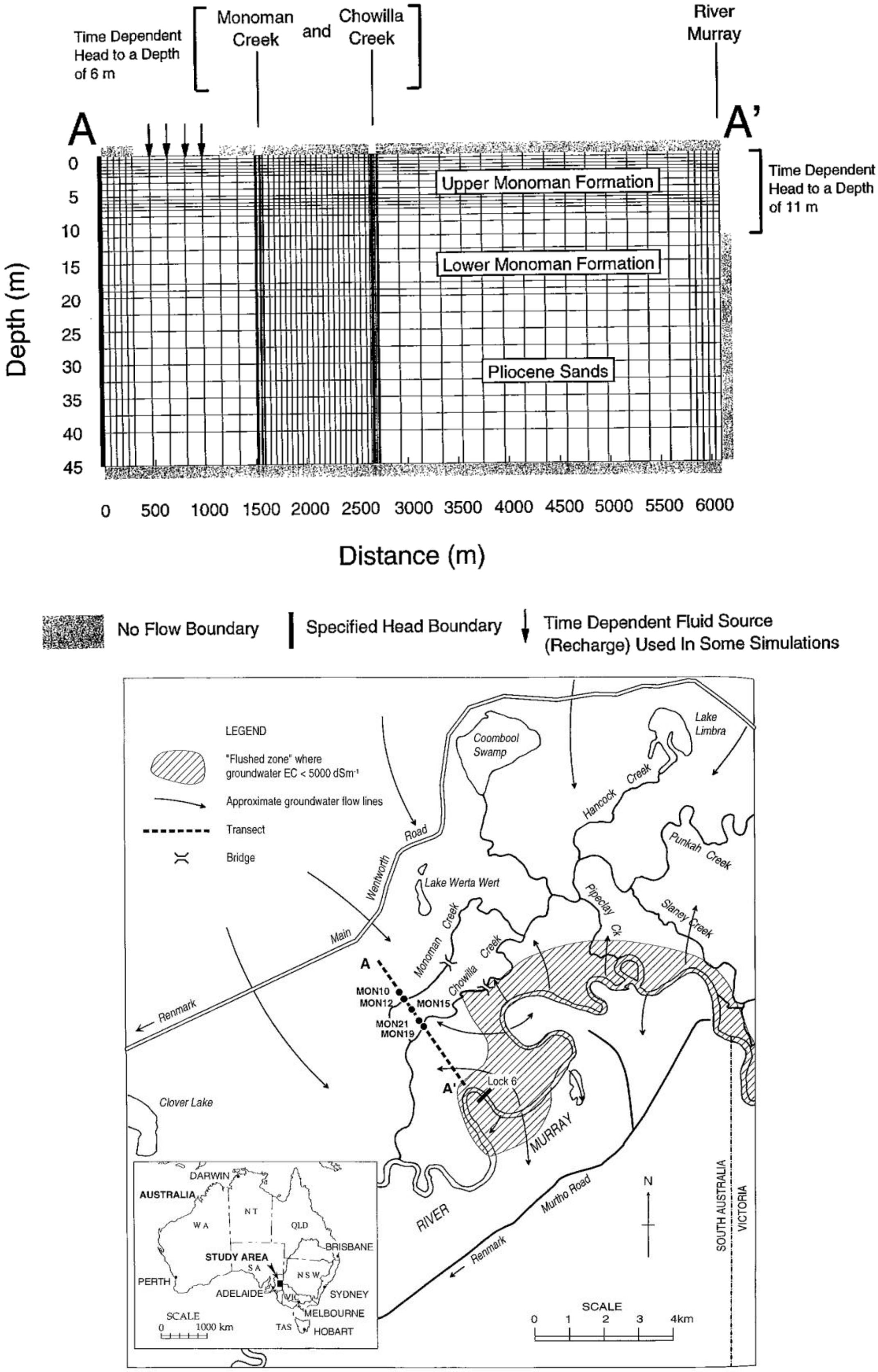


Figure 2-10 Model grid and peizometric transects for the cross-sectional Chowilla model (Source: Jolly *et al.,* 1998)

The simulated flood resulted in overbank flow and inundation recharge. Parameterisation of the clay linings around each of the creeks was the primary cause for poor matching (shape and timing) of the predicted and measured heads at the piezometers along the model transect. The initial simulations did not utilize SUTRA's ability to model saturated and unsaturated conditions simultaneously. Due to problems with calibration the model was re-run in the saturated-unsaturated mode with a new mesh. Considerably better matches to both heads and salinity were obtained when including the unsaturated zone processes.

The model showed that the greatest fluxes of water, and hence salt returning to the creeks from bank storage, occurred near the base of the creeks. This finding was also suggested by the field data. The results also showed that the Monoman Creek received greater salt loads from groundwater than Chowilla Creek which had been previously identified in an MDBC report. Bank storage mixing only accounts for early time salt recession to the creeks and that the addition of localised recharge on the floodplain is a plausible hypothesis to describe the later time salt recessions.

2.2 Groundwater model platforms

Physically-based numerical groundwater models of floodplains or floodplain processes in Australia have been developed predominantly with the industry standard groundwater model MODFLOW (Yan *et al.*, 2004; Doble *et al.*, 2005; RPS Aquaterra, 2012) and in some cases its solute transport counterpart MT3DMS (e.g. Murtho SIS climate sequence model). Other models have also been utilized on occasion such as SUTRA (Jolly *et al.*, 1998) and HydroGeoSphere (Doble *et al.*, 2011; Alaghmand *et al.,* 2014; Alaghmand *et al.,* 2015).

2.2.1 Software

MODFLOW is a 3D finite difference groundwater model developed by the United States Geological Survey (USGS) (McDonald & Harbaugh, 1988). It is considered the international standard for simulating and predicting groundwater conditions and groundwater/surface water interactions. Its modular structure allows it to be easily modified to adapt for a particular application. Its capabilities are expanded using related models, which includes MT3DMS for solute and SEAWAT for variable-density processes

SUTRA is a finite element model for saturated-unsaturated, variable-density groundwater flow with solute or energy transport in 2 or 3 dimensions (Voss & Provost, 2010). Simulation is through numerical simulation of a fluid and solute/energy mass balance equations. From literature reviews it appears to be not as commonly used at it was two decades ago for variable-density groundwater simulation. This is likely due to the increasing use of DHI's FEFLOW and the USGS SEAWAT (MODFLOW/MT3DMS based variable density simulator) by both research and consulting organisations. One major limitation is that SUTRA does not have adaptive time-stepping, which leads to difficulties in numerical convergence for some kinds of simulation.

HydroGeoSphere is a 3D control-volume finite element groundwater model capable of solving surface, sub-surface, solute and energy transport equation simultaneously (Therrien *et al.,* 2010). Surface water flow is modelled using a 2D depth-averaged flow equation which is the diffusion wave approximation of the Saint Venant equation for surface water flow. Groundwater is modelled using 3D transient Richards' Equation allowing for simultaneous saturated and unsaturated flow simulation. The model is capable of highly detailed fully-coupled surface and subsurface simulations and therefore lends itself well to floodplain process based modeling investigations. Implementing its vast array of features requires a greater number of input parameters requiring greater computational resources and ultimately longer simulation runtimes. Prior to 2013, the model was effectively public domain making it very popular among research institutions that were able to develop new versions of the model to suit specific numerical experiments. Its recent transition to a commercial software package will undoubtedly improve its capabilities but at the same time is likely to limit its uptake. One major limitation in its use for saline floodplains is that it does not simulate the evapoconcentration of salt; rather, it presumes that salt is removed as well as water during evapotranspiration.

Another option is MIKE SHE, which is a further development of the European Hydrological System-SHE. It is a deterministic, fully distributed and physically based modelling system for describing the major flow processes of the entire land phase of the hydrological cycle. MIKE SHE solves the partial differential equations for the processes of overland and channel flow and unsaturated and saturated subsurface flow. The system is completed by a description of the processes of snow melt, interception, and evapotranspiration. The flow equations are solved numerically using finite difference methods. In the horizontal plane the catchment is discretised into a network of grid squares. The river system is modelled with MIKE 11, which includes comprehensive facilities for modelling complex channel networks, lakes, reservoirs and river structures, such as gates, sluices, and weirs. MIKE SHE is applicable at spatial scales ranging from a single soil profile for evaluating crop water requirements to large regions comprising several river catchments. The groundwater simulation is more limited than MODFLOW.

Each of these platforms simulate different processes, with different options regarding process representation and complexity. As such, each has a different set of possible applications to modelling River Murray floodplain salinity dynamics. MODFLOW is sufficient where a model's primary aim is to estimate groundwater fluxes (or salt fluxes, if used in conjunction with the solute transport code MT3DMS). A variable-density couple flow-and-transport model, such as SUTRA, is necessary for the accurate simulation of features such as freshwater lenses, where there are significant variations in groundwater salinity. HydroGeoSphere is suitable for applications involving small or coarsely-discretised domains where simulation runtimes remain manageable; its fully-integrated approach is appealing, but the evapoconcentration error limits its application for saline floodplains.

MODFLOW with MT3DMS is selected as the groundwater modeling platform for the remainder of this study for reasons both conceptual and pragmatic. MODFLOW is one of the codes capable of simulating all the prioritised processes identified in Woods (2015a). Unlike HydroGeoSphere, the MODFLOW-MT3DMS combination can simulate evapoconcentration due to ET. MIKE SHE may also be a viable option, but the project aim is to assist DEWNR modeling capabilities. The timeframe of the project was short, and the need for DEWNR to adopt the project recommendations was immediate, so it is preferable to build on DEWNR's existing capabilities in MODFLOW rather than introduce a new groundwater modeling platform at this stage. That said, it is strongly recommended that other model platforms be investigated in later projects. A variable-density code will be needed to investigate the dynamics of freshwater lenses. Means should be explored for simulating interaction between the unsaturated zone and groundwater, starting with simple unsaturated zone modeling, then progressing in complexity. The use of fully-integrated models should also be explored.

2.2.2 Single model, multi model and integrated model approaches

Current approaches to modelling the lower River Murray and SA floodplains often involve the simulation of a single model simulating a single hydrological domain, without comparison to models of other domains. For example, the SA Salinity Register models of groundwater and the DEWNR/MDBA Source models of surface water are developed independently and their results are not compared to each other. For a region with considerable interaction between groundwater and surface water, one may also consider a multi model approach, where separate models of different hydrological domains are co-developed to be consistent with each other. Another option is to use a fully integrated model which simulates multiple domains simultaneously.

There are a growing number of hydrologic models that are capable of simulating integrated surface and sub-surface flow. Indeed, during the past 15 years, significant attention has focussed on the development and testing of physically-based integrated surface-subsurface hydrological models. Maxwell *et al.* (2015) recently reviewed several integrated models and also presented a benchmarking framework to diagnose their performance.

Despite the increasing research focus on integrated hydrologic models, single model and multi model management approaches are still currently presented in publications for regional, basin and catchment scale case studies. Sebben *et al.* (2013) attribute the lack of large scale applications of integrated models to the significant volume of data required to parameterize them. To the authors' knowledge, a definitive comparison of data, financial and time requirements for multi model versus integrated methods is yet to be published.

Multi model approaches generally comprise a suite of models each prioritising different physical processes. This approach may be considered fit for purpose when the following two conditions are met.

1. The dominant physical processes are well known and captured within the adopted suite of models and;
2. The fluxes within and between the models are in agreement and close the water balance.

The first point addresses the capabilities of the model directly. The latter point addresses the often overlooked aspect of identical internal water balance fluxes represented in more than 1 model of the suite. Ensuring these fluxes are in agreement is often not considered as it would require significant time investment for calibration, particularly if the models are developed and maintained by different departments or organisations.

Adopting an integrated model approach over a multi model is then highly predicated on the purpose of the model. If the purpose is to investigate effects of physical processes then the model's accuracy in representing those processes is paramount. If the purpose is for management where conceptual understanding is high then a model with a simplified representation of fluxes deemed less significant is likely adequate.

2.3 Modeling processes

The current floodplain conceptual model identifies numerous processes describing mobilisation to or from salt storage locations in the floodplain. For reasons given in Section 2.2 we focus our discussion on the various approaches used by the MODFLOW family of models in simulating the mobilisation processes. The conceptual model chapter of Woods (2015a) provides more detail on the processes.

2.3.1 Regional inflows

Regional groundwater flux to the floodplain occurs both laterally and vertically.

Lateral flux depends on the potentiometric head gradient at the boundary between the highland and floodplain. The flux will therefore vary over time depending on hydrogeological conditions in both the highland and the floodplain. Highland potentiometric head in the SA MDB depends on the regional hydrogeology (lateral and vertical flows within the regional aquifers, recharge from land use (native vegetation, land clearance, irrigation), and engineering works such as SIS.

Vertical flux depends on the potentiometric head gradient between the underlying regional aquifer and the floodplain aquifer; there is usually an intervening aquitard (an exception occurs west of Loxton, where the Monoman Formation lies immediately above the Pata Formation, which is a unit of the Murray Group aquifer system). Changes in the watertable elevation on the floodplain may result in changes to vertical flux between the Monoman and any underlying regional geologic unit. Increased flux to anabranches and isolated water bodies may also occur.

A floodplain groundwater model could specify either the potentiometric head or the flux at the highland-floodplain boundary. In MODFLOW, potentiometric head can be specified using constant head cells or general head boundary cells. Specified flux boundaries may be implemented using injection or extraction wells in each boundary cell. As the floodplain model will simulate changes in floodplain potentiometric head, it would be more accurate to specify a highland potentiometric head so that the gradient can vary over time.

Regional groundwater processes are simulated by DEWNR's SA Salinity Register models (Section 1.3). As such, the SA Salinity Register models could be used to provide boundary conditions for a floodplain model. Confidence in the predictive ability of the Salinity Register Models in SA implies that parameters for hydraulic and solute boundary conditions are readily available for the development of a site-specific floodplain model.

2.3.2 Evaporation and transpiration

Evapotranspiration is a major sink for groundwater on the floodplain and occurs in areas of shallow watertable (generally considered to be less than 3 to 4 metres below ground surface). Ensuring that accurate rates are used in models is imperative due to its prominent role in the water budget of floodplains along the lower River Murray.

The typical implementation of evaporation and transpiration in MODFLOW is via the ET package which lumps both processes into a single function. The approach is based on the following assumptions (1) when the watertable is at or above a specified elevation the loss from the watertable is at the maximum user specified rate (typically the land surface elevation in the model), (2) when the watertable exceeds a specified depth the rate is zero (watertable and capilliary fringe are too deep to be affected by negative pore pressure gradients created by plants roots and evaporation), (3) between these limits evapotranspiration varies linearly with depth.

A single function is not unreasonable in that the majority of cases both evaporation and transpiration result in a flux out of the modelled domain. The linear form of the function is restrictive and is not considered the best approximation of evapotranspiration behaviour for the majority of cases. Recent publications have identified that an exponential function better reflects the behaviour of a combined evaporation and transpiration depth dependent flux (Shah *et al.*, 2007). O'Grady *et al.* (2009) summarises published field derived transpiration values from a significant number of related studies focussing on floodplain vegetation.

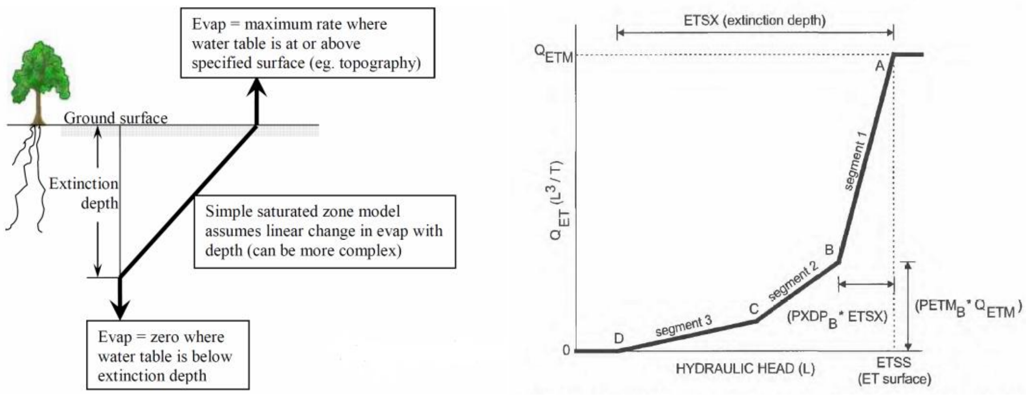


Figure 2-11 Comparison of the ET and ETS1 evapotranspiration functions (Source: Aquaterra)

An alternative ET package is available for MODFLOW, which permits segmented functions as opposed to linear namely the Evapotranspiration Segments (ETS1) package. Figure 2-11 illustrates the linear relationship of the standard ET package (left) and the segmented approach implemented in the ETS1 package (right). The logic behind multiple segments is that it allows the user to account for the variations in rooting density and in some cases the effects of contrasting soil horizons. With multiple segments the user can approximate an exponential depth dependent function for evapotranspiration however, it should be noted that functions with greater than 3 segments appear to cause instability (pers com. Noel Merrick). Alternatively, one could modify the source code of the ET package to change the linear relationship to that of an exponential one described in Bauer *et al.* (2006) in Section 6.2. Regardless of the package used, an accurate surface elevation data set is also essential as the rate of flux is governed by a combination of the function and modelled depth to watertable.

For well-defined areas of deep-rooted vegetation, the option of a negative recharge rate also exists. The logic behind this approach is that depth to the watertable does not play a significant role because plants are drawing water straight from the watertable. This approach is currently used in the Chowilla Model (RPS Aquaterra, 2012).

The best current approach for defining ET zones requires spatial information on vegetation class, vegetation health and groundwater salinity. ET rates are then assigned to areas of specific vegetation types based on literature derived maximum transpiration rates, transpiration salinity thresholds and extinction depths. Figure 2-12 gives an example of detailed ET zones as adopted in the EM4 model, which are based on a combination of groundwater levels, salinity, vegetation class maps and vegetation health. Maps showing ecohydrological types for the various floodplains provide a convenient dataset to delineate spatially variable ET rates.

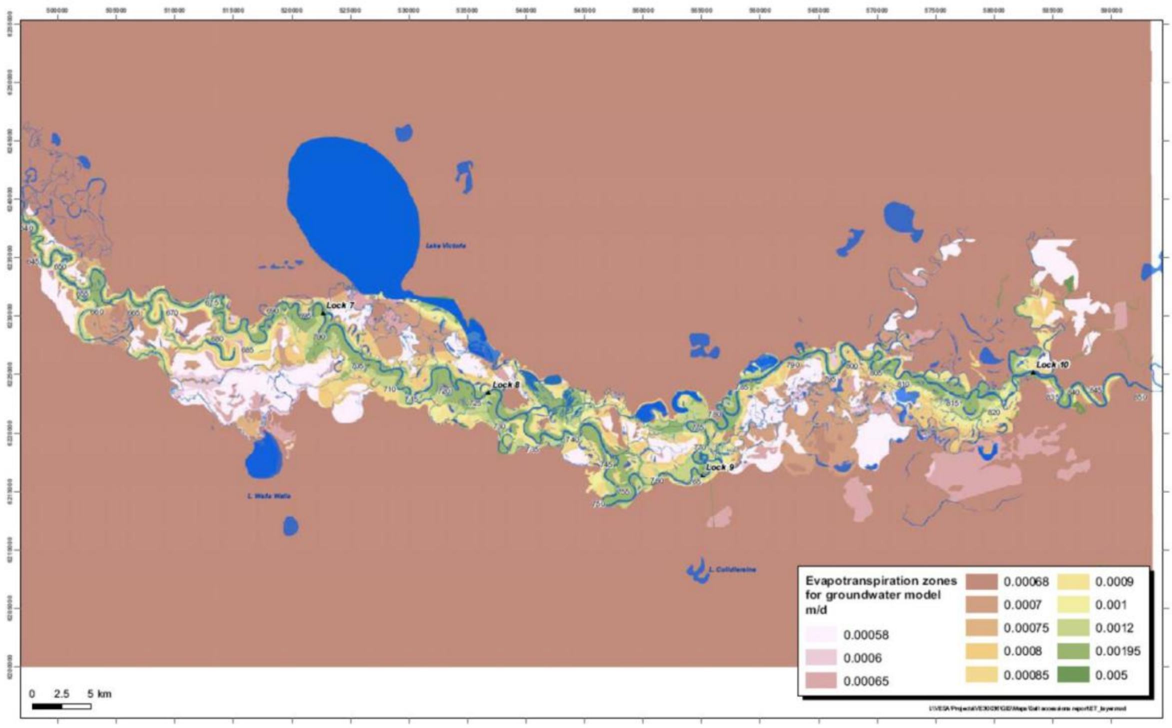


Figure 2-12 Evapotranspiration zones in the EM4 model (Source: Aquaterra, 2009)

A WAVES model may be used to improve accuracy of inflection points of the segmented functions assigned to specific vegetation/soil combinations. WAVES is an integrated one-dimensional unsaturated zone model that simulates energy, water, carbon, and solute balances of the soil-vegetation-atmosphere system on a daily time step. It also includes water and solute interactions between the soil zone and underlying groundwater. WAVES predicts the dynamic interactions and feedbacks between processes and hence is well-suited to investigations of vegetation growth response to climate and transient soil water and salinity status. The results from WAVES simulations would ensure that evapotranspiration fluxes are physically realistic for a specific location on the floodplain. The disadvantage to this method is the significant amount of data required as input parameters. WAVES has been adopted for numerous studies for floodplain vegetation. Slavich *et al.* (1999) used WAVES to develop a flood weighted root zone salinity index to assess vegetation health on the floodplain. Overton & Jolly (2004) used WAVES in a similar study focusing on vegetation health on the Chowilla Floodplain in response to flooding.

Ground-based direct measurements of evapotranspiration are also possible. Eddy covariance moisture flux towers are currently considered the most reliable method for measuring evapotranspiration over a footprint of hundreds to thousands of square meters (Drexler *et al.*, 2004). Ground-based measurements provide a reliable constraint for model derived evapotranspiration fluxes and if available should be considered integral to the model development process.

2.3.3 River level change

The River Murray and its changing levels can be simulated in various ways in MODFLOW. In the SA MDB, the river is always fully connected to the floodplain aquifer system, so it is not necessary to consider the unsaturated zone. In early versions of the Salinity Register models, the River Murray was simulated using constant head cells set at river pool level. In later versions, the constant head cells were replaced by river cells, so that the groundwater level under the river can differ from the river level itself. The river cells are assigned conductance values to represent the bed sediments. Conductance values are determined during model calibration and are varied as part of a sensitivity analysis. The river level is held constant in the Salinity Register models, except for the Chowilla model. Reservoir cells can also be used to represent the River Murray, as in the Murtho climate sequence model.

Reservoir cells provide greater accuracy as they interpolate the river level between the start and end of a stress period.

Significant level change in the Murray occurs due to flood and/or managed weir pool manipulation. Levels are generally monitored just upstream and downstream of a lock. The river level will change between two locks, following a curve that depends on the river flow rate. For natural floods, River Murray levels can be estimated from flow rates and published backwater curves, but there is often less information for anabranches. Where weirs are manipulated at the locks and/or regulators, river levels may be obtained from hydrodynamic models.

Most floodplain models have simulated river level change using stress periods of fixed length, typically monthly, which may not capture all the dynamics. The Murtho climate sequence model used stress periods based on the rate of change of the river level, permitting greater detail.

Changes in stage elevation also result in changes to the wetted perimeter of the channel. A representation of this effect within a MODFLOW model can be accomplished by also altering the flux-limiting river bed conductance of the river cells.

2.3.4 Inundation Recharge (includes flood and artificial watering)

The standard approach to recharge in MODFLOW is through the Recharge package, which simulates areally-distributed recharge to the groundwater system. Most commonly, areal recharge occurs as a result of precipitation and/or irrigation drainage that percolates to the groundwater system. The user needs to specify a value for recharge flux (in units of length per time) for each surface cell at each stress period. The entered value is multiplied by the horizontal surface area of the cell to produce a volumetric flux per time to be added (or in some cases subtracted) from the surficial aquifer. In the Lower Murray, precipitation recharge is negligible. Irrigation occurs on the floodplain at some locations, such as Lyrup and Renmark, but the flood inundation recharge represents the primary influx of water through the floodplain surface.

Representation of inundation recharge within a model requires knowledge of the timing and area of indundation and the rate of recharge. The time period of the inundation should influence the selection of model stress periods and time steps.

2.3.4.1 Area and timing of inundation

Different approaches have been used to determine the area and timing of floodplain inundation in the lower River Murray floodplain. The approaches include reservoir cells, aerial imagery (FIM), and hydrodynamic models, as described below.

The Murtho SIS climate sequence model (AWE, 2010) utilizes reservoir cells to simulate spatially-distributed recharge. Reservoir cells in MODFLOW were developed to simulate leakage to the underlying aquifer from a reservoir and behave in a similar manner to River cells in that they are essentially a head controlled flux boundary. The advantage of reservoir cells assigned to the elevation of the land surface is that it automatically activates when the reservoir stage is above the land surface. The user has to specify all model cells subject to inundation by each reservoir, the model layer number to which each active reservoir cell is connected, the land surface elevation of the reservoir bottom, the vertical hydraulic conductivity and thickness of the reservoir bed. Unlike the River cells, which have a fixed stage within a stress period the user has to prescribe starting and ending stage for the reservoir cell. The package uses linear interpolation to calculate the required stage elevation at each time step within a stress period. The reservoir bed conductance was used to represent the Coonambidgal Formations in the model. The surficial clay is often excluded from floodplain models of the Lower Murray due to the strong contrast in hydraulic conductivities between it and the underlying sandy Monoman aquifer, which leads to numerical instability.

CSIRO developed the River Murray Floodplain Inundation model (FIM), which provides commence to flow figures for different river flows and weir levels (Overton *et al.*, 2006). Figure 2-13 provides example output of the GIS FIM, showing an area around Lock 6 at Chowilla with flood predictions for 60,000 ML/day (dark blue) and 100,000 ML/day (light blue).

The following description of the FIM and its limitations is taken from Holland *et al.* (2005).

> "The FIM is a steady-state model that predicts the extent of flooding from a given flow on the first day of the flood. It does not consider the effect of antecedent conditions or the effect of flood duration. Further research on the wetting and drying behaviour of the floodplain and its wetlands needs to be incorporated into the model to be able to predict time sequences for management scenarios.The FIM is based on static remote sensing images taken at a range of river flows. As such, the areas of inundation that are derived from these images represent the flooding behavior under the floodplain and wetland management conditions at the time of the imagery. Lack of suitable imagery for River Murray flows above of 102,000 ML/d means that this is the maximum flow that FIM can model. The approximate floodplain elevations that can be derived from the model can therefore not be determined in areas which require a flow of greater than 102,000 ML/d for inundation."

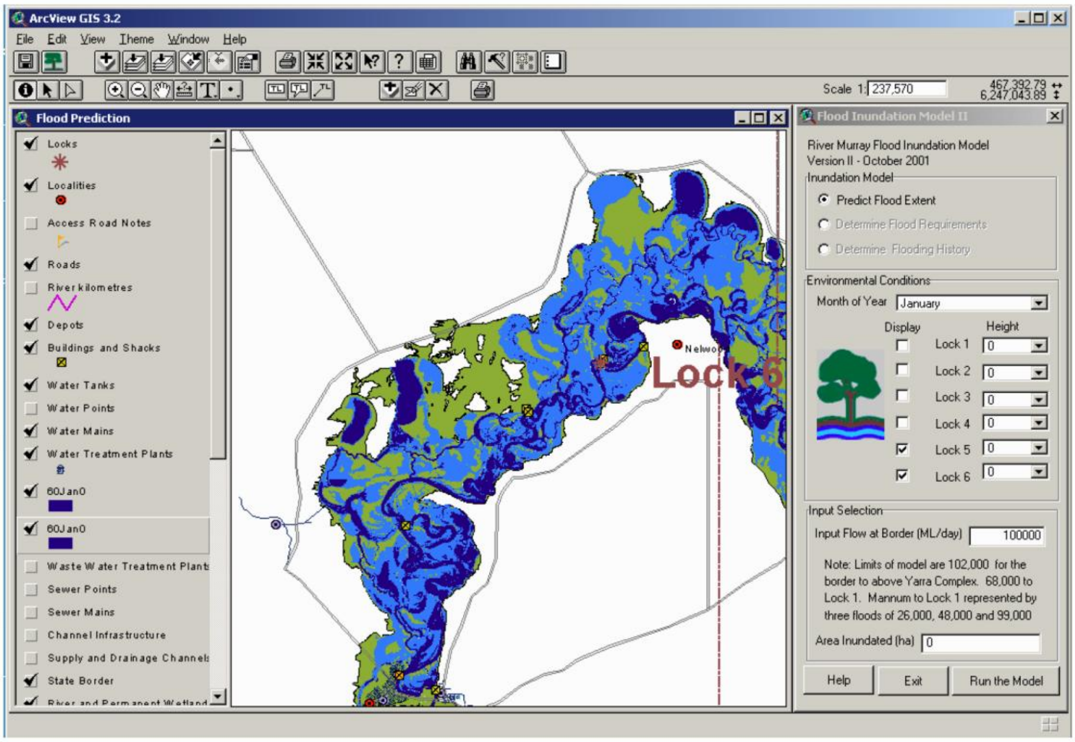


Figure 2-13 Example of FIM flood predictions

The third approach is a complete hydrodynamic model. Some exist for various reaches of the Lower Murray floodplain; however, some were developed by private companies and are not publically available. The primary advantage of the hydrodynamic model over that of the FIM is the transient change to inundated areas that may be obtained when simulating floods, increasing the accuracy of the timing of salt mobilisation processes. A hydrodynamic model may also make predictions for the impact of new infrastructure, such as regulators, which the FIM cannot. The model may be run at levels consistent with a hydrograph to establish detailed water level distributions in the anabranches, wetlands and inundation areas for groundwater recharge. Ensuring that the surface elevation map used in the hydrodynamic model is consistent with the groundwater model is imperative for accurate simulation of recharge.

The strengths and weaknesses of the three approaches vary. Hydrodynamic models are more sophisticated and are presumably more accurate but require significant data input and user expertise. They can simulate conditions that have not yet occurred, such as new regulators and levee banks. Where a hydrodynamic model is available, a groundwater model of a floodplain should use it. FIM estimates are available for the whole SA floodplain but are based on historical observations and so are not suitable for areas with new infrastructure. The reservoir cell approach calculates the inundation extent within MODFLOW, but does not account for details of surface water flow over the land surface such as sills.

2.3.4.2 Inundation recharge rate

The rate of recharge is the final component and arguably the most complex as many different approaches have been developed and used in various floodplain models. Field studies may reveal a prevalence of preferential pathways as cracks in the Coonambidgal Formation; however, these have been demonstrated to swell and shut as they wet through from inundation (Jarwal *et al.*, 1996). Recharge then becomes a function of vertical hydraulic conductivity, which can be estimated from soils maps. It should be noted that the soils of the lower River Murray are well understood, particularly the reaches between Locks 3 to 6, and have been described in detail in publications such as Jolly *et al.* (1993), Jolly *et al.* (1996a), Jolly *et al.* (1996b), Thorburn *et al.* (1992), Thorburn *et al.* (1995), Hollingsworth (1989), Jarwal *et al.* (1996) and Walker *et al.* (1994). The soil profiles are similar in nature, generally consisting of the Coonambidgal Formation overlying the Monoman Sand alluvial aquifer, with the major variation being thickness and clay content of the upper soil layer.

Overton *et al.* (2005) divided the Chowilla floodplain into three potential recharge zones with rates between 0.5 and 2 mm/day from soil parameters, remote sensing of vegetation response to flooding, and WAVES numerical experiments.

The EM4 model (Section 2.1.3) refines the spatial distribution further. Recharge estimates from soil data are combined with AEM slices showing resistivity at depth (Figure 2-14). It is presumed that where resistivity is low, highly saline groundwater is present, indicating less recharge. The range of recharge rates provided by the soils analysis is assigned to zones of indexed salinity with depth.

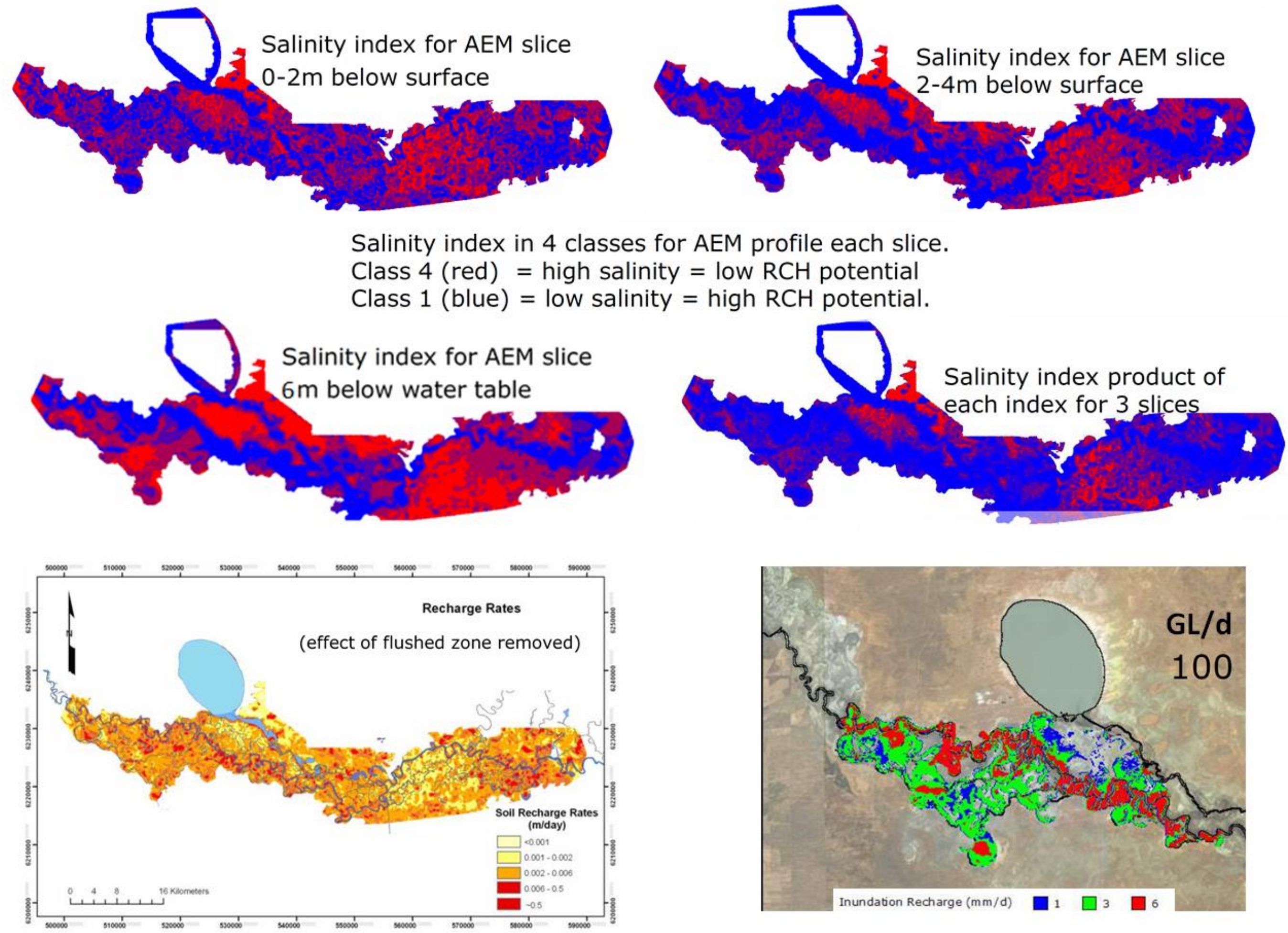


Figure 2-14 AEM slices and soil map used to produce the spatially-distributed recharge values for the EM4 model (Source: Aquaterra)

The reservoir cell approach (AWE, 2010) assumes that the inundated floodplain can be simulated as saturated flow. Recharge is calculated from the vertical gradient between the flood water level and the floodplain aquifer's potentiometric head, with a conductance value based on the vertical hydraulic conductivity of the Coonambidgal Formation.It is unclear whether these methods of estimating recharge produce significantly different rates, or how reliable they are. This should be investigated in Task 2 of this project.

Datasets comprising River Murray floodplain inundation recharge rates include those used in 1) the Floodplain Impacts model (Holland *et al.*, 2005), 2) the Bookpurnong Living Murray Pilot Project (Berens *et al.*, 2009), 3) Chowilla flooding study (Overton & Jolly, 2004). Bramley *et al.* (2003) also provides infiltration rates through the Coonambidgal via root channels.

2.3.4.3 Recharge salinity

Solute concentration associated with flood inundation recharge may require further investigation. It is currently unclear when modelling salt explicitly if neglecting leaching from the unsaturated zone is a valid assumption. Moreover, the salinity contrast may lead to significant density differences between water present in the floodplain and the fresher recharged water.

2.3.5 Evaporation from surface waters (includes interaction with wetlands)

This process is often handled by surface water or wetland-specific models however, it is possible to incorporate the effects into a MODFLOW model. Evaporation from surface waters leads to a loss in fluid and concentration of salts in the associated water body. For significantly large gaining surface water bodies that are disconnected from the River network, this may result in a constant flux of water exiting the floodplain at the maximum potential evaporation rate. A simple representation of this idealized scenario within MODFLOW would use extraction wells in each cell where the surface water body is located.

The MODFLOW Lake Package can simulate more complex interactions of surface water features with the floodplain aquifers. The following description from Merritt and Konikow (2000) describes the function of the Lake Package and its ability to account for evaporation:

> "In the Lake Package, a lake is represented as a volume of space within the model grid which consists of inactive cells extending downward from the upper surface of the grid. Active model grid cells bordering this space, representing the adjacent aquifer, exchange water with the lake at a rate determined by the relative heads and by conductances that are based on grid cell dimensions, hydraulic conductivities of the aquifer material, and user-specified leakance distributions that represent the resistance to flow through the material of the lakebed. Parts of the lake may become dry as upper layers of the model are dewatered, with a concomitant reduction in lake surface area, and may subsequently rewet when aquifer heads rise. An empirical approximation has been encoded to simulate the rewetting of a lake that becomes completely dry. The variations of lake stages are determined by independent water budgets computed for each lake in the model grid. This lake budget process makes the package a simulator of the response of lake stage to hydraulic stresses applied to the aquifer. Implementation of a lake water budget requires input of parameters including those representing the rate of lake atmospheric recharge and evaporation, overland runoff, and the rate of any direct withdrawal from, or augmentation of, the lake volume."

Unfortunately the solute transport integration of the Lake package is limited to MOC3D the method of characteristics solute transport model. Nevertheless, the hydraulic interactions of evaporating surface water bodies on the Lower Murray floodplains can be represented with this package given sufficient data.

2.3.6 Processes not modelled

Storage and release of salt in the unsaturated zone is generally not included in numerical models of the lower River Murray floodplain, but some field and numerical investigations have been done. Jolly *et al.* (1994) presented field observations of potentiometric head, hydrogeochemistry and soil salt storage before and after a flood. Soil salinity observations and modelling suggested that diffuse inundation recharge through the Coonambidgal Formation was minimal. They suggested that the sodic clays present in the Coonambidgal Formation swelled during inundation, resulting in very low infiltration rates, but that localised recharge could occur remote from the River, if the Coonambidgal Formation was not present. Akeroyd *et al.* (1998) also examined pre and post-flood soil salinity profiles at a different site and concluded the same. Charlesworth *et al.* (1994) examined the effect of diffuse versus localised recharge on the Chowilla floodplain with a 2D SUTRA model. Their model included solute and density effects. The results showed very little difference in solute flux to river between diffuse and localised recharge. Alaghmand *et al.* (2014) simulated salt storage in the unsaturated zone in a HydroGeoSphere model of Clark's Floodplain and found that 13% of the floodplain salt was stored there.

If it is deemed necessary to incorporate a representation of the unsaturated zone then the Unsaturated-Zone Flow package (UZF1) for MODFLOW may be used. The package uses a 1D kinematic wave approximation to Richards Equation which is solved by the method of characteristics simulating vertical flow only. The approach assumes that unsaturated flow occurs in response to gravity only and ignores negative potential gradients. It also assumes uniform hydraulic properties and uses the Brooks-Corey function to define the relationship between unsaturated hydraulic conductivity and water content. It effectively replaces both the ET package and the Recharge package. The UZF1 package differs from the Recharge package in that it adds water at the land surface and not directly to the watertable. Infiltration is limited by the saturated hydraulic conductivity. It also differs from the ET package because losses are first removed from the unsaturated zone and then if demand is not met the water is removed directly from groundwater when the level is less than the extinction depth. The package also allows for water to be discharged directly to land surface when the altitude of the watertable exceeds land surface elevation. Water discharged to land surface as well as infiltration excess may be routed as inflow directly to streams or lakes if those packages are active. A recently developed HYDRUS package for MODFLOW is implemented in a similar fashion to the UZF package and also offers better solute transport options.

Fluid density may alter groundwater flow and solute transport. The density of groundwater in SA MDB floodplains depends on salinity; their large variation in groundwater salinities means a large variation in fluid density. This is likely to be important for situations such as simulating a freshwater lens. In these situations, a groundwater code should be used which can simulate variable-density groundwater flow and transport, e.g. SUTRA, HYDRUS or SEAWAT. A notable paper specific to freshwater lens behaviour in the SAMDB is Massmann et al. (2006) whose study identified that a critical salt concentration difference exists, which maximises the salt discharge to a surface water body. A density invariant approach to modelling freshwater lenses will not account for this if large contrasts between surface water and groundwater salt concentrations are present.

2.4 Data requirements and availability for floodplain groundwater modelling

Woods (2015a) provides a list of datasets relevant to SA River Murray floodplain dynamics. Table 2-1 lists the information required for a detailed groundwater model of a floodplain in the study area.

Table 2-1 Data requirements for a floodplain groundwater model

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Construction	Ground surface	Lidar	Ideal if accurate to the cell size used at the surface in a model. Important for ET which is often controlled by depth to water. Upscaling/downscaling may be an issue.	Available throughout floodplain at good resolution.	River Murray DEM, DEM
Construction	Thickness/extent of Coonambidgal	Scout holes, borelogs	Often excluded from floodplain models due to numerical issues; however, could be critical for numerical experiments of flood recharge.	Few floodplain bores in most areas. Could revist AEM to interpret.	Waterconnect DB, SA Geodata DB, SARIGDB?, SA Soils DB
Construction	Thickness of Monoman	Borelogs	Essential for providing an accurate transmissivity of the watertable aquifer in the floodplain. May also indicate if water table is within the Coonambidgal, which has ramifications for ET. Ideally accurate to the cell size of the surface layer of a model.	Few floodplain bores in most areas. Could revist AEM to interpret.	Waterconnect DB, SA Geodata DB, SARIGDB?
Construction	Units below Monoman	Borelogs	Essential for indicating the presence of an aquitard below the Monoman. Has a significant effect on the flow dynamic (ie. lateral vs. vertical) below the Monoman. Most critical west of Loxton, where the Booxton aquitard may not be present	Few floodplain bores in most areas.	Waterconnect DB, SA Geodata DB, SARIGDB? SIS DB

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Construction	Lateral regional boundary conditions, gw flow	Potentiometric head in Loxton Sands near the floodplain	Used to inform boundary conditions at the edge of a floodplain model also useful as calibration in the regional Salinity Register Models. Critical for estimating regional groundwater flux into the floodplain.	Reasonable level of interpolated information in most areas.	Waterconnect DB, SA Geodata DB, SARIGDB?, SIS DB, Salinity Register Models
Construction	Lateral regional boundary conditions, solute transport	Salinity in Loxton Sands near the floodplain	Essential if modeling salt explicitly or including density effects. Can also be used as a calibration target.	Reasonable level of interpolated information in most areas.	Waterconnect DB, SA Geodata DB, SARIGDB?, SIS DB, Salinity Register Models
Construction	Vertical regional boundary conditions, gw flow	Potentiometric head in the Pata Formation under the floodplain	Will indicate whether vertical or lateral flux dominates. Can also be used as a calibration target.	Very few floodplain bores in most areas.	Waterconnect DB, SA Geodata DB, SARIG DB?, SIS DB, Salinity Register Models, Calperum bores
Construction	Vertical regional boundary conditions, salinity	Salinity in the Pata Formation under the floodplain	Essential if modeling salt explicitly or including density effects. Can also be used as a calibration target.	Very few floodplain bores in most areas.	Waterconnect DB, SA Geodata DB, SARIGDB?, SIS DB, Calperum bores
Construction	Recharge, rainfall rate	Monthly rainfall records	Recharge component of a groundwater model.	Available	BoM

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Construction	Recharge, irrigation rate and area	Irrigation history on floodplain	Could have a significant effect on salinity in the floodplain. Will need volumes and area of irrigation	Available; may require refinement in some areas	SIS DB
Construction	Recharge zones, inundation	AEM and/or EM	Can be used to inform recharge zones through the Coonambidgal, critical for estimating impacts of floodplain recharge.	Available for some portions of the floodplain only	AEM, Calperum AEM
Construction	Recharge zones, inundation	Vegetation type and health	Necessary for spatial variation in ET. May also be used as a surrogate for floodplain salinity in the absence of salinity data	Last complete review was 2007; needs to be updated since the Millenium Drought and subsequent flood	BDBSA DB, Ecohydrological_Types
Construction	Recharge rates, inundation	Salinity data before and after a flood	Essential for informing the effect of the Coonambidgal on the floodplain. May also be useful in ground truthing the use of AEM to infer spatial variation in recharge.	Minimal data	Literature
Construction	Inundation areas, historic	Floodplain Inundation Model estimates (FIM)	Simulating the effect of natural floods. Steady state snap shots only.	Available; results could be compared with the more detailed hydrodynamic models available for some regions	RiM-FIM III, FIRU

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Construction	ET rate, potential	Monthly potential ET from BoM	Essential for evapoartion from surface water bodies and ET	Available	BoM, CMRSET
Construction	ET extinction depth	Root depth?	Possible from spatial distribution of vegetation types/classes	Needs to be developed	BDBSA DB, Ecohydrological_Types
Construction	ET rate curve with depth	Ecological literature?	Will inform the function used to describe ET behaviour. Most recent literatue suggests that an exponential function be used	Needs to be developed	Literature
Construction	River Murray levels, historic	River levels at Locks	Informs heads for internal boundary conditions for floodplain models. Ideal temporal scale monthly or less.	Available	Surface Water Data DB, Hydstra DB
Construction	River Murray levels, historic	Backwater curves	Informs heads for internal boundary conditions for floodplain models.	Available for the main channel	Surface Water Data DB, Hydstra DB

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Construction	River Murray bed conductance	Heat tracers? Other technique?	Essential for internal boundary conditions where ground and surface water interact.	Minimal data	Literature, Calperum Flush zones
Construction	Anabranch levels, historic	Any observations?	Informs heads for internal boundary conditions for floodplain models.	Minimal data	Surface Water Data DB, Hydstra DB (maybe)
Construction	Anabranch bed conductance	Heat tracers? Other technique?	Essential for internal boundary conditions where ground and surface water interact.	Minimal data	Literature
Construction	Coonambidgal hydraulic conductivity	Laboratory tests?	Necessary if modeling the Coonambidgal explicitly. Also necessary if incorporating its effect into ET or flood inundation models.	Minimal data; some for Bookpurnong and Chowilla	Literature
Construction	Monoman hydraulic conductivity	Aquifer tests	Necessary as a model input parameter. Could be calibrated as well.	Few aquifer test results	Waterconnect DB, SA Geodata DB, SARIGDB?, SIS DB

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Construction	Monoman hydraulic conductivity	Texture from borelogs	Texture indicates the heterogeneity	Available for a few locations only	SIS DB
Construction	Coonambidgal porosity	Laboratory tests?	Necessary as a model input parameter. Need to know if there is change to the porosity when inundated. Will cause a significant change to the hydraulics on the floodplain if it does.	Minimal data	SA Soils DB, Literature
Construction	Monoman specific yield	Aquifer tests	Necessary as a model input parameter.	Few aquifer test results	Waterconnect DB, SA Geodata DB, SARIGDB, SIS DB
Construction	Monoman porosity	Laboratory tests?	Necessary for solute transport simulation	Minimal data	SA Soils DB; Literature
Construction	Monoman storativity	Aquifer tests	Essential groundwater model input parameter.	Few aquifer test results	SIS DB; Literature

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Construction	Underlying aquitard vertical hydraulic conductivity	Aquifer tests	Essential groundwater model input parameter. May be a calibrated parameter. Has a significant effect on the flow dynamic (i.e. lateral vs. vertical) in the floodplain. Best if known to within an order of magnitude.	Few aquifer test results	SIS DB; Literature
Construction	Dispersivity, Monoman	Textbook values?	Necessary for solute transport simulation	Minimal data	Literature
Construction	Salinity, groundwater	Salinographs from the Monoman Formation	Necessary for solute transport simulation	Few floodplain bores in most areas.	Waterconnect DB
Construction	Salinity, groundwater	Salinity with depth, Monoman Formation	Can be used as a calibration target in a floodplain salt model. Also can be used to calibrate an AEM survey.	Minimal data	Literature
Construction	Salinity, groundwater	AEM and/or EM	Useful when modeling salt on the floodplain explicitly. Could be used for initial conditions depending on timing of survey	Available for some portions of the floodplain only	AEM, Calperum AEM

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Construction	Salinity, initial	AEM	Could be used for initial conditions to provide spatial variation to the salt concentration on the floodplain depends on timing of survey	Available for some portions of the floodplain only	AEM, Calperum AEM
Construction	SIS	Locations and pumping rates	Essential input for groundwater models due to significant effect on flux to the floodplain and river	Available; seek regular updates from SA Water	SIS DB
Construction	Salinity, groundwater	Texture, salinity, moisture in upper soil profiles, depth to water	Can be used as a calibration target in a floodplain salt model. Also can be used to calibrate an AEM survey.	Available across SA floodplain; CSIRO smb://144.110.3.110/clw-share9/floodpl/Floodplain_SA_field_results	FIP Soil Profiles
Calibration	Potentiometric head, Monoman	Hydrographs	Useful groundwater models as both calibration targets and initial conditions	Few floodplain bores in most areas.	Waterconnect DB, Surface Water Data DB, Hydstra
Calibration	ET rate	Satellite data of ET, spatial, monthly	Large component of the floodplain water balance so spatial variability and accuracy important.	Coarse data available; finer resolution should be developed	CMRSET

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Calibration	ET rate	Field observations	Eddy Covariance Flux Towers are best used in conjunction with satellite-derived ET. Other field based measurements may be useful to observe daily/seasonal variation.	Available 2010-current	Calperum Flux Tower
Calibration	ET rate	Water balance plots, soil profiles, water potentials	Can be used as a calibration target in a floodplain salt model. Also can be used to calibrate an AEM survey.	Available; 2005, 2007-2008	Bookpurnong Floodplain Pilot Project
Calibration	Gaining/losing river	NanoTEM, main river channel	Useful for indicating sections of a reach that are gaining and/or losing. Can be used to check hydraulics in a numerical model. May also be indicative of the bed hydraulic conductivity.	Available for 2004 and 2012	NanoTEM
Calibration	Gaining/losing anabranches	NanoTEM, anabranches	Useful for indicating sections of a reach that are gaining and/or losing. Can be used to check hydraulics in a numerical model. May also be indicative of the bed hydraulic conductivity.	Available for some anabranches	NanoTEM
Calibration	Salt load to River Murray	Run of River	Effectively base flow conditions and may indicate reaches that are significant salt contributors.	Available	Salinity Register Models DB

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Calibration	Salt load to River Murray and anabranches	Toroidal coil salinity	Useful for indicating salinity contributions from anabranches. Also used to partition flood salt load contributions from specific floodplains (AWE & CMC 2012)	Available in recent years	Waterconnect DB
Calibration	Salinity distribution	AEM and/or EM	See Salinity, groundwater	Available	AEM, Calperum AEM
Calibration	Salinity distribution	Groundwater salinity with depth	See Salinity, groundwater	Available	AEM, Calperum AEM
Calibration	Salinity distribution	Changes in vegetation health	Indicates where has salinity has changed; part of calibration confirmation	Available (2002, 2007, 2009 surveys)	TreeHealth, TreeHealth2007Katarapko, TreeHealth2009RiverlandRamsar
Calibration	Recharge distribution	Geochemical estimates of recharge	Useful for freshwater lens history and for constraining model recharge	Part of ARC Linkage study of freshwater lenses	Waterconnect DB? Literature.

Model phase	Model feature	Data	Usage	Availability	Goyder Data Review Entry
Scenarios	River Murray levels, future	Hydrodynamic model estimates	Hydrodynamic models provide better temporal resolution for river stage levels and inundated area which may affect floodplain hydraulics and inundation recharge	Available for Chowilla, Pike and Katfish	RiM-FIM III, FIRU
Scenarios	Anabranch levels, future	Hydrodynamic model estimates		Available for Chowilla, Pike and Katfish	RiM-FIM III, FIRU
Scenarios	Inundation areas, future	Hydrodynamic model estimates		Available for Chowilla, Pike and Katfish	RiM-FIM III, FIRU
Scenarios	SIS	Locations and pumping rates/target levels	Essential input for groundwater models due to significant effect on flux to the floodplain and river	Available when SIS designed, SIS bores and total pump volumes for 2011/12/13/14	SIS DB, SA_SIS_Interim16012014
Scenarios	GCS	Locations and pumping rates/target levels	Essential input for groundwater models due to significant effect on flux to the floodplain and river	Available when GCS designed	

2.5 Discussion

2.5.1 Monitoring recommendations

Where a detailed model is required of an area's floodplain salinity dynamics, the modelling should be supported by a monitoring program. The scope of the program will depend on the aims of the project, budget and time. Below is a list of monitoring tasks which would inform the modelling. For many of these items, there will be existing data and infrastructure which should be reviewed before a monitoring plan is developed.

The monitoring program should consider:

- Scout holes and bore logs to provide: thickness of the Coonambidgal Formation, thickness of the Monoman Formation, thickness of the underlying aquitard (if present)
- Observation wells in the Monoman Formation, monitoring potentiometric head and salinity
  - some wells constructed so that salinity can be observed at different depths within the aquifer
- Observation wells in the aquifer below the Monoman Formation monitoring potentiometric head, with some initial sampling of salinity
- Aquifer tests using the observation wells to determine the aquifer properties of the Monoman Formation and the vertical conductivity (leakage) of the underlying aquitard
- Observation wells in the regional watertable aquifer, near the floodplain edge, monitoring potentiometric head, with some initial sampling of salinity
- Aerial and/or land-based electromagnetic surveys, as these indicate the distribution of groundwater salinity, both laterally and with depth
- In-river transient electromagnetic (RTEM, also known as NanoTEM) geophysics surveys of anabranches to determine where groundwater salt is entering the river
- Monitoring of water and salinity levels in anabranches and wetlands to quantify storage and release of salt
- Surveys of vegetation type and health, as these indicate both current and past groundwater conditions. For example, healthy trees will be found where they have access to freshwater from the surface water system or groundwater, while dying trees indicate a change for the worse in access to freshwater.
- Collate leaf area index (LAI) data; this can capture the nature of vegetation (type and health) and also soil type (different vegetation types prefer different soils) and is derived from NDVI or LiDAR remote sensing data
- Gather data from weather stations based on the floodplain (particularly those on or near water) monitoring rainfall and ET; these can be used to ground-truth remotely sensed estimates of AET (e.g. CMRSET)
- Laboratory and field analysis of the hydraulic properties of the Coonambidgal Formation
- Consider whether an eddy covariance tower is required to groundtruth CMRSET satellite-derived estimates of AET
- Field estimates of riverbed, anabranch and wetland conductance. Expert opinion should be sought as to the most appropriate method.

For areas where a detailed model needs to be calibrated against the results of a trial (e.g. artificial/environmental watering trial or GCS trial, or possibly a natural flood), include the following at locations where changes in soil and/or groundwater salinity are anticipated, before and after the trial:

- Aerial, land-based and/or in-river electromagnetic surveys
- soil salinity sampling
- vegetation type and health survey
- continuous logging of groundwater levels in key bores

To improve understanding of salinity dynamics in the study area, we recommend the following:

- A review of observation well locations and the frequency of potentiometric head and salinity monitoring.
- A review of existing AEM data, to identify and prioritise floodplain areas where: (i) no AEM data currently exists, or (ii) significant changes in salinity are expected to have occurred since the last AEM survey. Priority should be given to areas where new management works (such as regulators) are proposed.

2.5.2 Groundwater modelling priorities

The primary issues are of modelling method, scale (temporal and spatial), and data requirements. The priority processes have been identified as follows:

1. **Bank storage/river level change with solute transport.** Investigate the:
    a. Impact of stress period length on model outputs, particularly groundwater salt load to the River.
    b. Impact of using reservoir cells rather than river cells, due to their different interpolation over time of river levels
    c. Impact of fully simulating solute transport on salt load estimates, as compared to using flux calculations multiplied by fixed salinity values
    d. Impact of grid size
2. **Evapotranspiration.**
    a. Compare the methods of representing ET as used in standard Salinity Register models, the Chowilla model, the EM4 model and the Shahse River Valley model.
    b. Compare the salt load estimates made when ET is constant versus when ET varies seasonally.
    c. Develop recommendations for modelling ET which depend on the desired level of accuracy and data availability.
3. **Floodplain inundation recharge.**
    a. Compare the methods of representing inundation recharge used in the Chowilla model, the EM4 model and the Murtho SIS climate sequence model.
    b. Determine impact of spatial and temporal discretisation
    c. Develop recommendations for modelling inundation recharge which depend on the desired level of accuracy and data availability.

4. **Density-dependence.** Develop a cross-sectional model to investigate the importance of density in simulating the interface between fresh river water and saline groundwater, i.e. the extent of a freshwater lens. The model will require a code capable of simulating variable-density flow and solute transport.

The first three processes are investigated using numerical models in Chapter 4. The fourth process is being investigated in the ARC-Linkage project, *Dynamics and management of riverine freshwater lenses,* a collaboration between Flinders University, Monash University and DEWNR.

3 Surface water modeling approaches

Peter Cook

3.1 Introduction

There are a very large number of models that have been constructed for simulating river flow, and the interaction between rivers and their floodplains. These models can be divided into two broad groups – hydrodynamic models and hydrologic models. In general, hydrodynamic models solve differential equations for conservation of mass and momentum. They model flow in terms of physical characteristics of the river (such as roughness coefficients). Hydrologic models use a routing approach based on rating curves to simulate flow using a conceptual reservoir approach.

This chapter gives a general overview of some of the most common surface water models (with an emphasis on those models that have been applied in Australia), and a description of their functionality for representing wetland and river floodplain systems.

3.2 Hydrodynamic Models

There are a large number of hydrodynamic river models, many of which have been used in Australia. For predicting inundation of river floodplains, three different approaches are possible. In increasing order of complexity, these are:

1. Quasi 2D Approaches based on 1D Hydrodynamic Modelling
2. Combination of 1D Hydrodynamic Model with a 2D Floodplain Reservoir Model
3. Full 2D Hydrodynamic Modelling

3.2.1 1D and Quasi 2D Hydrodynamic Approaches

The one-dimensional Saint-Venant equation for shallow surface water flow can be written:

$$S_f = S_o - \frac{\partial h}{\partial x} - \frac{u}{g}\frac{\partial u}{\partial x} - \frac{1}{g}\frac{\partial u}{\partial t} \tag{3-1}$$

where S_f is the friction slope, S_0 is bed slope, u is velocity, g is gravitational acceleration, x is distance and t is time. The terms on the right hand side of Equation 3-1 represent bedslope, local inertia (or acceleration), advective inertia, and pressure differential terms. Often simplifications to these equations are used that neglect different terms to reduce computational complexity. Thus the diffusive wave approximation neglects the last two terms, and the kinematic wave approximation neglects the last term.

Many river flow models rely on the one-dimensional solution to the Saint-Venant equation, or one of its simpler approximations. Both HEC-RAS and MIKE 11 are examples of such models, and solve the full Saint Venant equations for river flow using implicit finite difference schemes. To model the floodplains, a quasi two-dimensional approach can be applied, in which the flooded areas are modelled as separate 1D river branches (Figure 3-1). These are connected to the main river using spills or overflow structures that represent the embankment elevations between the river bed and the floodplains. Thus, the floodplains are modelled by a network of fictitious river branches and hydraulic structures. The details of the branching systems used by HEC-RAS and MIKE11 are discussed in more detail by Villazon & Willems (2009). The approach requires that the pathways across the floodplain are defined by the modeller in advance. Cross-sections perpendicular to these flowpaths are derived from a digital elevation model (DEM), and roughness coefficients can be chosen to represent the floodplain vegetation. Huang *et al.* (2007) describe the use of this approach using the DYNHYD hydrodynamic model for simulating flooding due to breaching of dykes over a nine day period for a 50 km reach of the Elbe River, Germany.

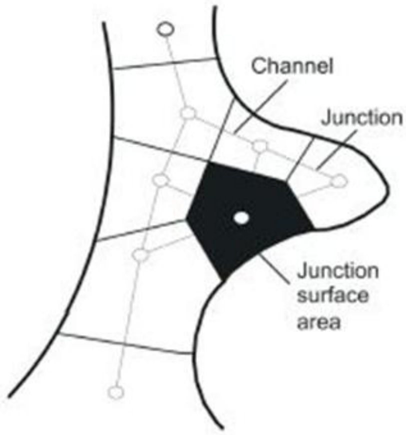


Figure 3-1 Quasi-2D approach to floodplain modelling, whereby the floodplain is simulated as branches of the main river channel (Source: Huang *et al.*, 2007).

3.2.2 1D Hydrodynamic River Model and 2D Floodplain Reservoir Model

A number of studies use a 1D hydrodynamic model for the main river channel, and a reservoir approach for the river floodplain. One of the simpler such approaches is the model developed by Paiva *et al.* (2011). This approach uses the full Saint Venant equations for 1D river flow and a simple storage model for the floodplain units. If the water level in the river exceeds the height of the levee bank, then it spills over and inundates the floodplain. Floodplain units are identified between river cross-sections, and water flows between the river and the floodplain but cannot flow between adjacent floodplain units. Thus the water level for each floodplain unit is equal to the water level in the main channel adjacent to the floodplain unit. Each floodplain unit is thus characterised by simple water depth – flooded area relationship, and is hydrologically connected to a river segment. The model was applied to the Purus River basin, one of the subbasins of the Amazon River. Floodplain width in Purus River basin is approximately 30 km, and approximately 130 floodplain units were defined in the model (Figure 3-2). This is a simple approach which focuses on water exchange between the river and the floodplain, rather than water movement within the floodplain itself. The model thus does not represent floodplain hydrodynamics, but can describe temporal patterns of floodplain inundation and the impact of floodplains on river hydrographs.

Other models use a grid-based approach for the floodplain. The advantage of this approach is that the pathway of the flood on the floodplain is simulated in the model, and does not need to be pre-defined. One such model is LISFLOOD-FP (Bates & De Roo, 2000; Wilson *et al.*, 2007), which represents channel flow using the kinematic wave approximation to the full 1D St Venant Equations, and floodplain flows using a reservoir approach implemented on a raster grid. The model assumes that flow between floodplain cells, and between the river and floodplain cells, is simply a function of the difference in free surface water height between those cells. Thus the dimension, elevation and friction coefficient for each floodplain cell is defined. Each cell is treated as a storage volume, and water can flow in any direction across the floodplain according to the water surface gradient and the local topography. Manning's Equation is used to calculate the flow rate between floodplain cells in each time step of the model, and the roughness parameter can be adjusted depending upon the type of vegetation cover on the floodplain.

Bates & De Roo (2000) present an example of the application of the LISFLOOD-FP model to the River Meuse, on the border between The Netherlands and Belgium. The model covered a 35 km reach of the river, and had 9,639 nodes and 18,939 elements, and was run for a 21 day period to capture inundation from large flood. The model described by Bates & De Roo (200) did not include direct precipitation onto the floodplain or runoff from upland areas, or floodplain losses due to evapotranspiration and infiltration, although these could be readily added.

Subsequently, Wilson *et al.* (2007) presented an application of the LISFLOOD-FP model to a 13,000 km^2 section of the central Amazon floodplain, Brazil.

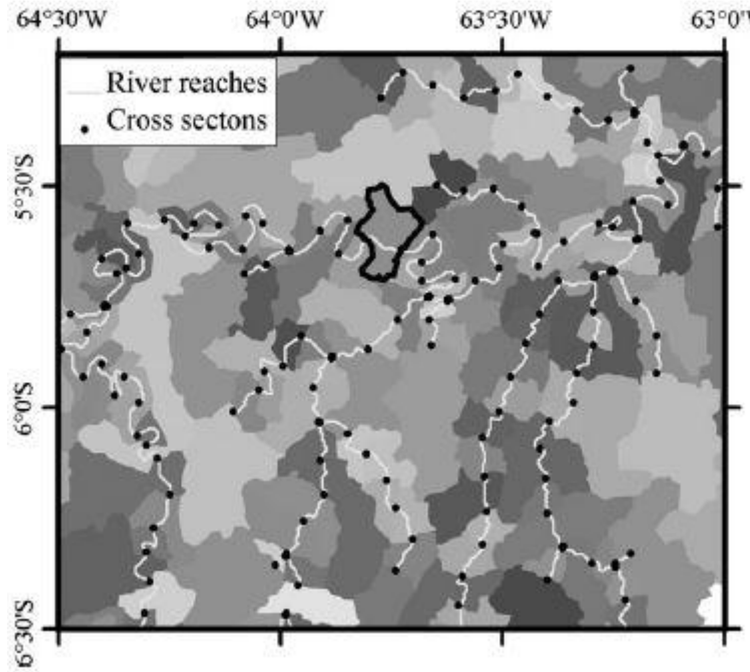


Figure 3-2 Location of river reaches, cross-section and floodplain units (grey shaded regions) on the Purus River Basin, Brazil (Source: Paiva *et al.*, 2011).

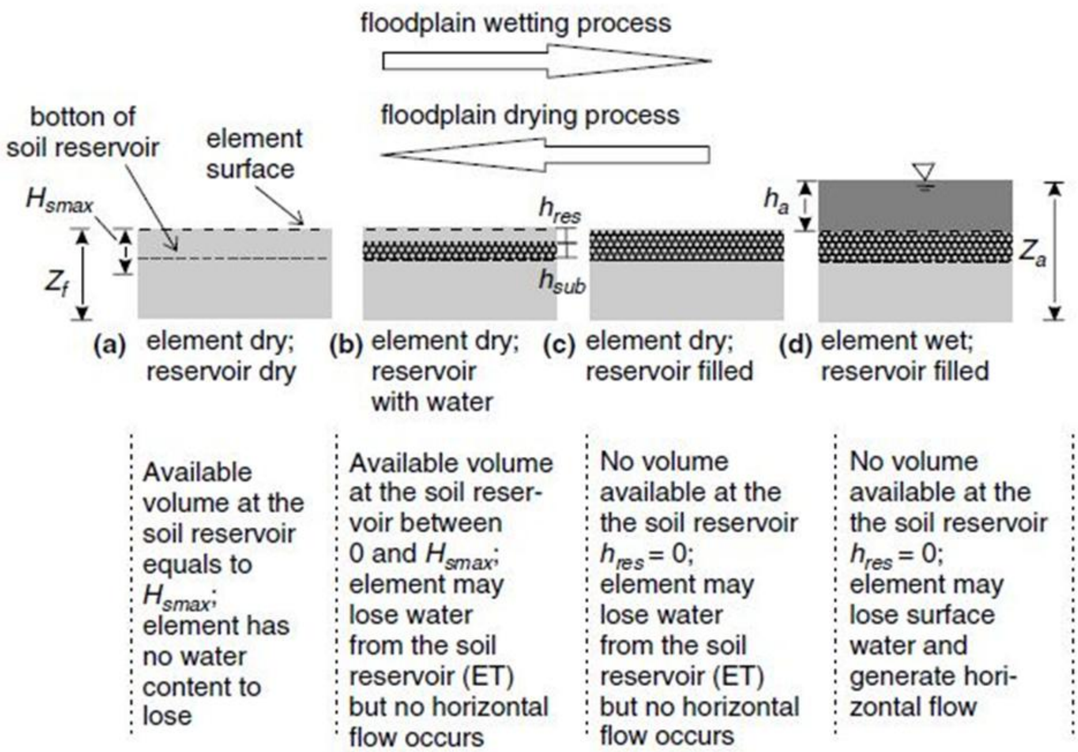


Figure 3-3 Wetting and drying process of floodplain elements of the raster model of Paz *et al.* (2011). Z_f is floodplain elevation, Z_a is water level, h_a is surface water depth, h_{sub} is water depth of soil reservoir, h_{res} is the available volume of soil reservoir, which has a maximum capacity equal to H_{smax}.

Paz *et al.* (2011) use a similar approach to examine floodplain inundation in the Upper Paraguay River Basin. The authors coupled the 1D hydrodynamic model IPH4 with a floodplain raster model based on the LISFLOOD-FP approach. However, the authors approach differed from the LISFLOOD-FP model in that flow between floodplain elements was based upon flow through channels, whose width and length could be specified and was not related to the floodplain element grid size. The model also included a vertical water balance for the floodplain cells. Surface water only occurred on the floodplain when the soil moisture storage was exceeded. Input to the soil moisture storage occurred by flow from adjacent floodplain units and as precipitation. Loss occurs by surface flows and by evapotranspiration, with the evapotranspiration rate dependent upon the soil moisture storage (Figure 3-3).

3.2.3 Full 2D Hydrodynamic Approaches

River flows across floodplains can also be simulated using full two-dimensional solution to the river hydrodynamic equations. MIKE 21 is one such program. Variables are defined across a grid, and the model solves the full time-dependent non-linear equations of continuity and conservation of momentum. MIKE 21 can be used for simulation of water flow in rivers, lakes, estuaries and coastal areas, as well as across floodplains. Infiltration on the floodplain can also be simulated using the MIKE 21 HD Infiltration and Leakage module. The MIKE FLOOD model dynamically couples the one-dimensional MIKE11 river model and the two-dimensional MIKE21 model by defining linkages between the two models that allow water to move between the 1D and 2D domains.

Critical data for the use of such models is a digital elevation model with sufficient horizontal and vertical resolution to capture the important topographic features of the floodplain. Model calibration is usually achieved by adjusting channel and floodplain roughness coefficients, and sometimes also adjusting cross-section and topographic data, to reproduce recorded hydrographs and observations of floodplain inundation (Wen *et al.*, 2013). The main advantage of the 2D approach is that it provides velocities and discharges in 2 dimensions. The path of the floodwave is determined from topographic information, and does not need to be specified a priori.

Tuteja & Shaikh (2009) used the MIKE FLOOD model to simulate surface flows across the Koondrook-Perricoota Forest to support development of environmental watering plans and engineering design of structural works, and also compared results from the MIKE FLOOD model (which combines 1D and 2D approaches) with results of Quasi-2D and 2D approaches. The Koondrook-Perricoota Forest covers an area of 337 km^2, and is located north of the River Murray floodplain between Torrumbarry Weir and Barham. The 2D MIKE 21 model used a grid resolution of 40 m. However, a comprehensive MIKE 11 model was developed at a fine resolution to capture inundation dynamics and flow exchange between channels and the floodplains, which could not be adequately represented by the MIKE 21 floodplain model alone. Thus six floodplain channels were delineated, using 809 cross-sections up to 200 m width and described at 1 m intervals. The models were validated using remote sensing mapping of inundation areas for 12 images across three historical events (1991, 1993 and 2000), and also mapped flood extents for the 1946 flood. The authors noted that without including the floodplain channels, the 2D floodplain MIKE 21 model overestimated inundation extent, whereas the Quasi 2D MIKE 11 model underestimated inundation extent, as observed in satellite imagery and historical records. The authors also observed that inundation extent had a hysteretic relationship with river flow (Figure 3-4), which reflects water trapped on the floodplain following peak river flows. The hysteretic relationship was not predicted by the Quasi 2D MIKE 11 model.

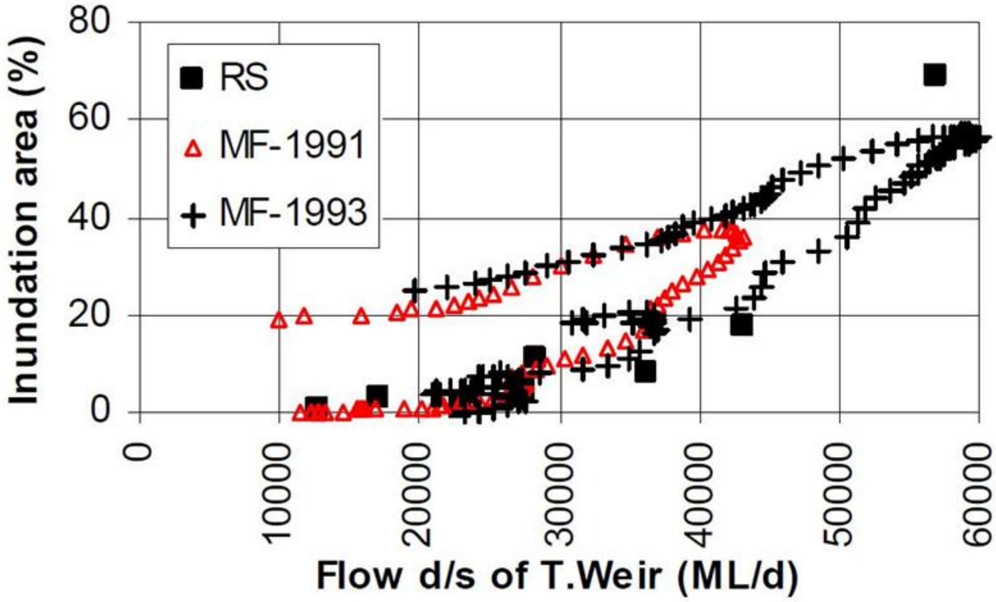


Figure 3-4 Hysteretic relationship between river flow rate and floodplain inundation area, observed for the Koondrook Perricoota Forest. Inundation area is higher for the same river flow rate on the falling limb of the flood (Source: Tuteja & Shaikh, 2009).

Chen *et al.* (2012) used the MIKE FLOOD model to simulate a 500 km^2 portion of the Florida Everglades using 400 m x 400 m grid cell, to simulate both flow and solute transport using the conservative chloride ion. They noted that although the 400 m grid resolution was the best topographic resolution available, it was still too coarse to capture some of the topographic features that control shallow water flows. They also noted that the ability to define depth-dependent roughness is important for chloride transport simulations, and this that capability is not currently available in MIKE21.

3.2.4 Hydrologic Models

River reservoir models use a hydrologic routing approach to calculate flow. Most of these models consider the river as a series of nodes connected by links. Nodes are points on the river system that have certain operation or physical processes associated with them. These include gauges, tributary of pumped inflow locations, on-river storages, diversion points and pumps, wetlands and floodplain detention storages. Water flow and water quality routing occurs along links which separate the nodes (Simons *et al.*, 1996).

Hydrologic routing relies on the continuity equation. In its simplest form, inflow to the river reach is equal to the outflow of the reach plus the change in storage.

$$I(t) = O(t) + \frac{\partial S}{\partial t} \tag{3-2}$$

where *I* is the inflow to the reach during, *O* is outflow from the reach; and *S* is the water storage in the reach.

Reservoir models have the advantage of small computational times, and so they are suitable for inclusion in decision support systems, and for optimisations and uncertainty analysis, which require many model runs. This type of model is widely used for river management in Australia, and examples include BIGMOD, REALM, IQQM and Source.

The MDBA uses MSM-BIGMOD to model river flow and salinity. The model (Close, 1996; MDBC, 1996, 2002) is specific to the River Murray system, which is represented as a number of river reaches and branches. MSM is a monthly simulation model used for modelling flows, operating rules and irrigation demands, and BIGMOD is a daily flow routing and salinity model that extends from Hume Dam to the barrages at the Murray Mouth. BIGMOD uses a water balance approach, with the downstream flow of each reach calculated as:

$$q_{out} = q_{in} + \Delta S - d - E \times A - L_{hf} - L_{em} \tag{3-3}$$

where q_{out} is the flow out of the reach, q_{in} is flow into the reach, ΔS is the change in reach storage, d is the distributed diversion, E is the net evaporation rate, A is the reach surface area, L_{hf} is the high flow losses and L_{cm} is the continuous monthly losses. The storage in each reach is a function of the flow rate, which is contained with a lookup table. A second lookup table contains the relationship between flow rate and travel time for each reach (Rassam and Werner, 2008).

BIGMOD has been widely used for modelling salinity within the River Murray, and salinity within the model is calculated based on the water balance. Salinity values are assigned to different particles of water, and their movement is tracked through the model. Lakes are assumed to be fully mixed, and weirs may either be assumed to be fully mixed, or can be simulated as part of the river. Salt loads from tributaries and drains are included in the model, and increases in salinity can also occur due to evaporation, with decreases in salinity due to dilution from precipitation or inflow of water having lower salinity. The model minimises the difference between simulated and observed salt concentration in the river by introducing an unaccounted salt inflow. (Negative values indicates salt outflow.) Unaccounted salt inflow accounts for processes that are not explicitly simulated in the model (including groundwater inflows), as well as errors introduced by the various model assumptions and simplifications. Telfer *et al.* (2012) describe application of the BIGMOD hydrologic model to simulate salinity within the lower River Murray system. The model simulates both flow and salinity, with excellent fits to observation data (Figure 3-5). However, very high unaccounted salt inflows are required, representing 35% of the total salt loads to the river reach from Euston to Morgan over the 39 year period from July 1970 to June 2009. The unaccounted salt inflows are thought to be derived almost entirely from floodplain sources (Telfer *et al.*, 2012).

REALM and IQQM are river hydrology models developed to simulate the operation and optimisation of water supply systems. REALM is widely used in Victoria and supported by the Victorian Department of Sustainability and Environment (DSE), wheras IQQM is preferred in New South Wales. Both models use a node-link approach, and are able to represent a wide range of river features and operating rules. Mass balance calculations are applied for off-river storages and wetlands. In particular, IQQM simulates the floodplain as a series of reservoirs (or cells), each with its own specific depth – area – volume relationship. A water balance is performed for each wetland, with change in storage equal to the difference between inflows and outflows. Inflows occur due to surface water flows from adjacent wetlands or from the river, and outflows occur due to evaporation and surface water flows to adjacent areas. A storage can have inflows and outflows from and to multiple locations, to reflect the complexity of floodplain topography. Direction of flow on the floodplain can change depending on the size and nature of inundation extent. In the case of two adjacent cells, flow direction can vary depending upon which cell gets inundated first.

Source simulates river systems in a similar manner to IQQM, but includes some additional functionality to represent some simple surface water – groundwater interactions (Welsh *et al.*, 2012). In particular, the impacts of groundwater pumping and recharge on river – aquifer exchange are included. Simple analytical expressions are used to estimate the time delay between groundwater pumping and flow to the river, based on the distance of the bore to the river, and the aquifer diffusivity (Rassam *et al.*, 2009). However, the model does not simulate salt transport directly. Rather, the salinity is defined by the user, and the model calculates salt loads to the river by multiplying the groundwater flow rate by the specified salinity.

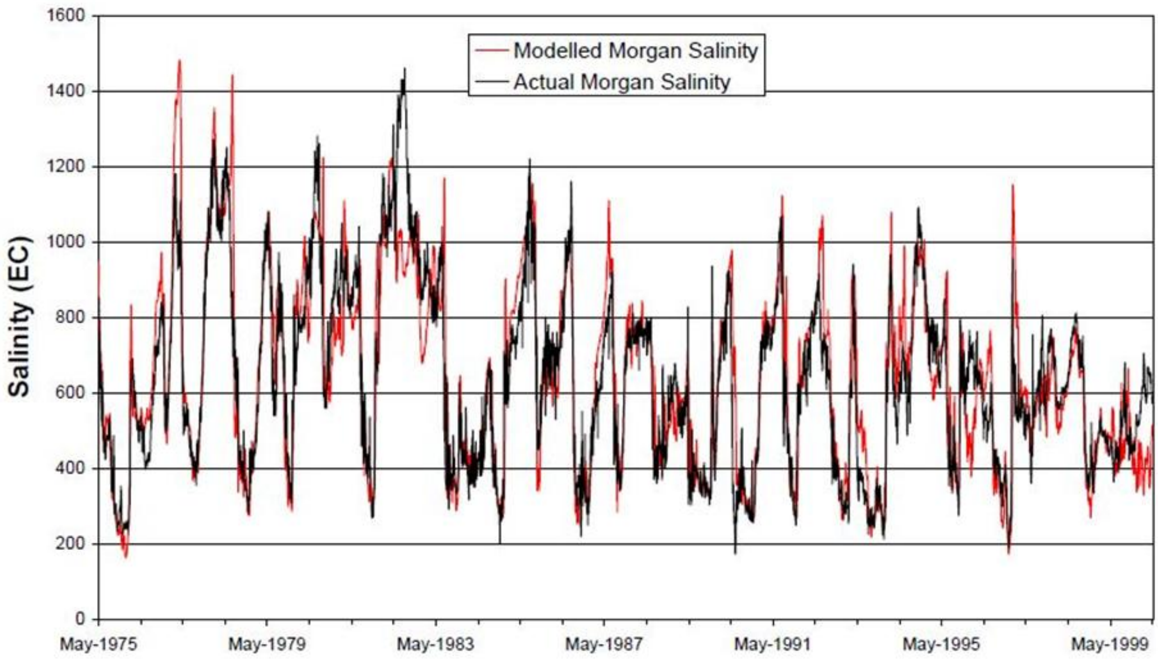


Figure 3-5 Comparison of BIGMOD simulated salinity at Morgan with observation data for the period 1975 – 1999 (Source: Telfer *et al.,* 2012).

3.3 Combining Hydrodynamic and Hydrologic Models

A number of studies have attempted to combine the more accurate simulation of hydrodynamic models with the run-time advantage of hydrologic models. The principle of this approach is to use the hydrodynamic models to develop relationships between flow and inundation, and to embed this relationship into a hydrologic model. For example, Mackay *et al.* (2011) used the MIKE 21 2D hydrodynamic model to determine the relationship between flow and inundation for a section of the Lower Murrumbidgee River centred on Yanga National Park, near Balranald, NSW, and used this relationship in an IQQM model. The hydrodynamic model involved a grid size of 20 m to simulate an area of approximately 300 km^2. (Comparisons were also made with a hydrodynamic model using a coarser 80 m grid cell; Figure 3-7) Based on results of the hydrodynamic simulations, the floodplain area was split up into 38 cells, each with its own water level – discharge relationship (derived from the hydrodynamic model). Water level – discharge relationships were derived for boundaries between all reservoirs, and between the reservoir and the river. Many cells had multiple inflow and outflow points, reflecting the complexity of the floodplain topography. Direction of flow on the floodplain can change depending on the size and nature of inundation extent. In the case of two adjacent cells, flow direction can vary depending upon which cell gets inundated first. Following calibration, the results of the IQQM model were compared with inundation extent mapped using satellite data obtained at 84 discrete time intervals within the period 1988-2006. Although there were strong correlations between mapped and observed inundation extent, the modelled inundation extent was consistently lower than observed extent (Figure 3-6). It was found that modelled results better matched observed results for larger floods than for smaller flood events. (The modelled inundation area is averages 79% of the observed inundation area in wet years, but only 44% during drier periods.) Areas that were poorly simulated included those that received flows from other areas of the floodplain, rather than directly from the river.

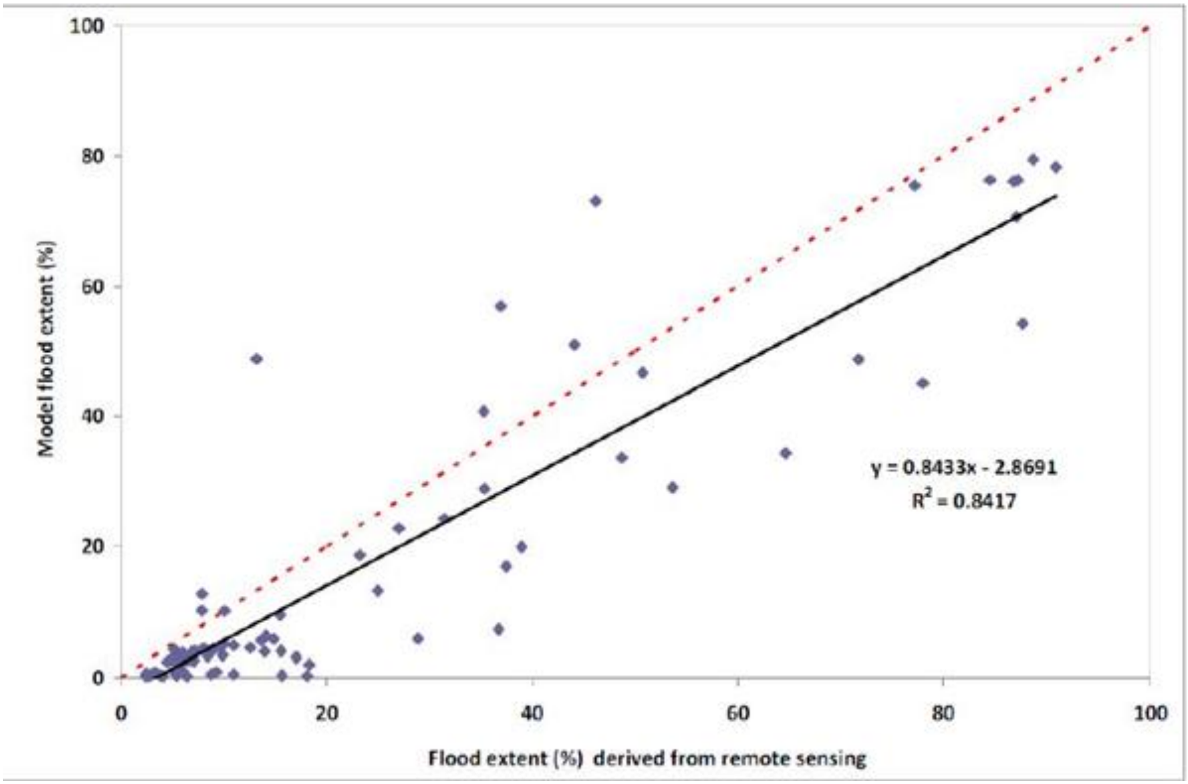


Figure 3-6 Relationship between modelled and observed inundation extent in a section of the Murrumbidgee River floodplain, Yanga National Park, NSW (Source: Mackay *et al.,* 2011).

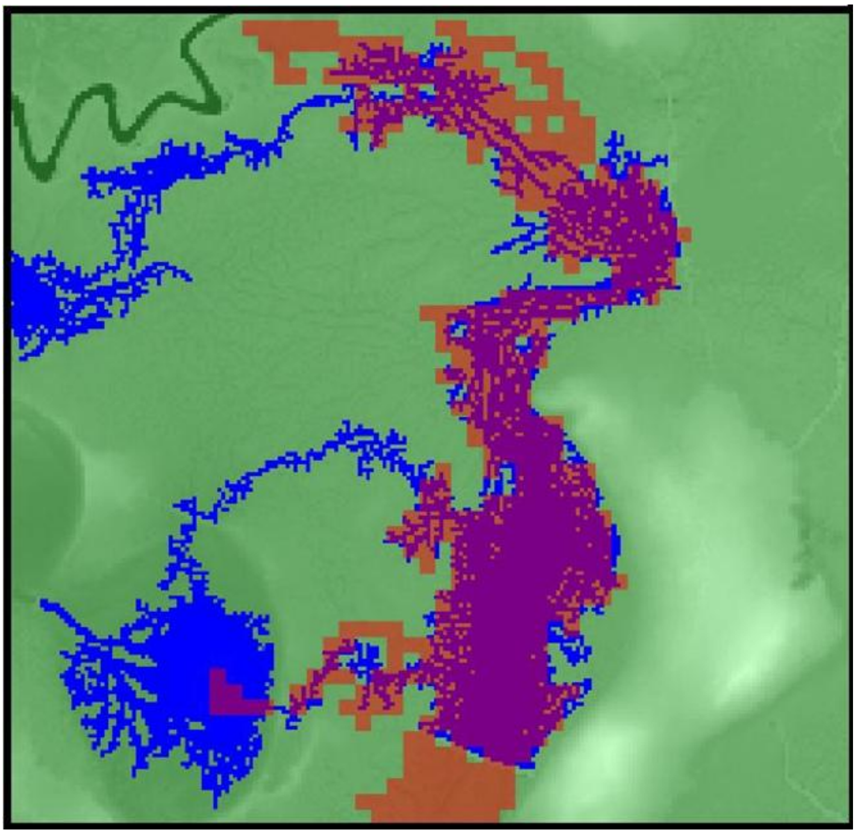


Figure 3-7 Comparison of the effect of grid cell size on simulation of inundation extent of the Murrumbidgee River in Yanga National Park, NSW using a hydrodynamic model. Simulated floodplain inundation is much greater using a 20 m grid cell (blue areas) than using a coarser 80 m grid (red and

purple areas). (Purple areas are inundated in both models, and green areas are uninundated.) (Source: Mackay *et al.,* 2011).

A similar approach was used by Wen *et al.* (2013). These authors developed a spatially distributed 2D hydrodynamic model for the Macquarie Marshes using the MIKE FLOOD modelling system, and then used this to prepare IQQM hydrologic model. The hydrodynamic model included almost 500 km of river channels (simulated in MIKE 11), and 213443 active cells across the floodplain (simulated in MIKE 21). The MIKE FLOOD model was calibrated against six historic flow events, covering the full range of flow conditions in the marshes. The complex relationships between flow and inundation apparent from the hydrodynamic model were then simulated with an IQQM model with 36 storages to represent key wetlands, and a total of 88 inflow pathways to these wetlands. In IQQM the inflow to each wetland is determined by the river level (or level of another wetland), and this relationship was derived from the MIKE FLOOD model. The model was used to simulate effect of water extraction for irrigation on marsh inundation.

3.4 Discussion

The choice of simulation model should be based on an assessment of the key processes that need to be represented to produce accurate predictions. However, it is also often influenced by practical considerations, such as model run time. Wen *et al.* (2013) note that long model run times is the main impediment to the widespread use of 1D and 2D hydrodynamic models for river and floodplain simulation. Such models often take hours to days for single simulations, depending on the size of the modelled area, and time period of the simulation and the spatial and temporal discretisation. For example, Chen *et al.* (2012) report that MIKE FLOOD modelling of the Florida Everglades took 70 minutes for one year simulation (on 2.83 GHz Intel® Core(TM)2 Quad CPU), while the LISFLOOD-FP hydrodynamic model of the Amazon wetlands described by Wilson *et al.* (2007) took 14 days to run (3.0 GHz PC). Tuteja and Shaikh (2009) report simulation times of 15-30 minutes for quasi 2D MIKE 11 modelling, 5 – 15 days for 2D MIKE 21 modelling and up to 50 days for combined 1D and 2D MIKE FLOOD modelling of the Koondrook Perricoota Forest wetlands (Intel Dual-Core PC with 2.13 GHz processor speed and 3.25 GB RAM). These longer run times can become prohibitive if models are to be incorporated into decision support systems or used for uncertainty analysis, both of which can require large numbers of model runs. Run time can be reduced by increasing model cell size, although this can affect model results. If the grid cells are too large, the model will not accurately simulate smaller floods that inundate only parts of the floodplain. Reservoir models have much faster run times which allow them to be used in decision support systems and for assessment of multiple management options. Thus Mackay *et al.* (2011) report an IQQM runtime for 100 year simulation at daily time scale of only a few minutes. The longer run times of the hydrodynamic models also means that it is more difficult to apply these at larger spatial scales, and this is the area where the simpler hydrologic models have their main advantage.

When using simple models, it is important to understand the assumptions behind them, and consequent errors that can arise when these assumptions do not hold. The historic driver for many of the hydrologic models (e.g., IQQM, REALM) has been the need to manage water allocation and water use associated with irrigation and town supply (Hameed and Podger, 2001). They were not explicitly designed for management of environmental flows, and so wetlands were either not represented, or represented so coarsely as to be of little value for simulating floodplain processes (Wen *et al.*, 2013). However, these models are now being used to simulate floodplain inundation, and so their accuracy for this purpose needs to be assessed. Examination of the BIGMOD simulation of lower River Murray salinity reveals that a large fraction of the salt loads into the model are unaccounted salt inflows, which are essentially a statistical calibration parameter to represent processes that are not explicitly simulated by the model. The model would therefore not be capable of simulating changes to the processes affecting this salt inflow.

The Source model has recently been modified to include changes in groundwater flow to the river due to irrigation (an increase in groundwater flow to the river) or groundwater pumping (which results in a decrease in groundwater flow to the river). Salt loads are calculated by multiplying the groundwater inflow to the river by a salinity, which is specified by the user. Thus changes in salt loads to the river due to groundwater pumping or irrigation are assumed to be proportional to the change in groundwater flow rate. The model does not simulate salt transport directly, and so cannot simulate changes in the concentration of groundwater discharging to the river. However, it is likely that temporal changes in concentration of groundwater discharging to the river occur due to processes within the floodplain, although the magnitude and significance of these is not yet clear. Thus flooding of floodplain areas may lead to infiltration of relatively fresh groundwater, which would ultimately discharge to the river. In between floods, evaporation within the floodplain environment would concentrate salt, increasing concentrations in the groundwater. Thus the concentration of groundwater discharging to the river may vary in time depending on the flooding history. None of the current models can simulate this process, even in a rudimentary form. Although many of the available hydrologic and hydrodynamic surface water models are also able to simulate water quality (and in particular, salinity), and a few contain simple groundwater flow representations, none of them are able to simulate leaching of salts beneath the base of floodplains or wetlands, and the subsequent transport of that salt back into the river. Of course, the importance of these concentration changes is currently unclear, and analysis of the significance of these processes would be valuable.

Comparison between predictions of simple models and more complex models should precede the decision to use simple models, but this important step is frequently overlooked. Simple models simplify or omit certain flow or transport processes, and will be in error if these processes are important. Simpler models also have lower data requirements than complex models. However, even if the data required by the complex models is not available, it is still possible to carry out sensitivity analysis for these unknown parameters using the complex models. This allows the effect of the unknown (or poorly known) parameters on model predictions to be determined. This is not possible using simple models, and so the effect of simplifying assumptions often cannot be explicitly determined.

The data required to construct accurate hydrodynamic models of river floodplain and wetland systems includes surface elevations on a fine resolution across the entire floodplain area, and information on hydraulic properties (e.g., surface roughness). Within recent years, the availability of accurate digital elevation models has increased, so that surface elevation data is now often available with a spatial resolution of a few tens of metres and vertical accuracy of 1 – 2 m. Higher resolution is available using LiDAR systems. Surface roughness data is often obtained using literature values that relate it to vegetation type and density, but also used as a calibration parameter. Hydrologic models determine inundation and flooding using volume – area relationships and relationships between river and wetland level and flow rates between river and wetland (and between individual wetlands or floodplain elements). Spatially explicit information on precipitation, evaporation and infiltration for the floodplain is also required where these components are included in the models. However, observations of water level and discharge and spatially infrequent – flow gauges are usually restricted to the main river channel, and even water level data is often absent across large floodplain areas. Remote sensing observations of inundation extent do not completely fill this void, and they are often limited in spatial resolution and temporal coverage (Schumann *et al.*, 2009). One option is to develop a 2D hydrodynamic model with high spatial resolution to simulate inundation, and to use the relationships derived from this for hydrodynamic models. When this approach is used, the calibration of the hydrologic models can only be guaranteed for the range of river flood levels simulated using the hydrodynamic model from which it was derived.

4 Groundwater model trials

Virginia Riches, Tariq Laattoe, Juliette Woods & Carl Purczel

The aim is to simulate key floodplain processes and drivers using MODFLOW in a simplified model of the SA River Murray floodplain. Numerical experiments are performed to explore how process representation affects model outputs. As such, the model is a "generic model" according to the model classification discussion of the *National Water Commission Groundwater Modelling Guidelines* (Barnett *et al.*, 2012):

> "These models are developed primarily to understand flow processes and not to provide quantitative outcomes for any particular aquifer or physical location. They can be considered to provide a high level of confidence as their accuracy is only limited by the ability of the governing equations to replicate the physical processes of interest. While they provide high confidence when applied in a general, non-specific sense, if the results are applied to or assumed to represent a specific site the confidence level will automatically decrease."

Generic models are a special case and cannot be easily matched to the Confidence Level Classification of Barnett *et al.* (2012).

The primary issues are of modelling method, scale (temporal and spatial), and data requirements. Both groundwater flow and solute transport are simulated. Base case models are developed, simulating a variety of floodplain conditions where processes and drivers are constant over time. River level, ET, and groundwater recharge are homogeneous and do not vary over time. The base case models are calibrated to match reasonable ranges of groundwater flux to the floodplain, groundwater salt load to the river, and actual ET.

Once the base case models are satisfactory, they are modified to simulate scenarios in which drivers vary over time. Different parameters and process representations are trialled. The priority processes are:

1. **River level change.** A MODFLOW/MT3DMS model tests the following, without simulating floodplain inundation:
    a. Impact of temporal discretisation of changes in river level on model outputs, particularly groundwater salt load to the River.
    b. Impact of using reservoir cells rather than river cells, due to their different interpolation over time of river levels.
    c. Impact of fully simulating solute transport on salt load estimates, as compared to using flux calculations multiplied by fixed salinity values.
    d. Impact of grid size.
2. **Evapotranspiration.** Develop a MODFLOW model to investigate the following:
    a. Impact of temporal discretisation of changes in PET on model outputs, particularly ET flux and river conditions.
    b. Sensitivity to ET parameters.
    c. Impact of spatial variations in ET.
    d. Impact of the shape of the ET function with respect to depth to water.
3. **Floodplain inundation recharge.** Develop a MODFLOW model to do the following:
    a. Compare the impact of floods with and without inundation.
    b. Consider the impact of a wetland (backwater) on floodplain dynamics

The results of the numerical experiments are used to develop recommendations for the simulation of floodplain salinity processes in the SA MDB.

4.1 Model Conceptualisation

The Goyder Floodplain Model has been designed to represent generic conditions for the SA River Murray floodplain in the study area (Figure 4-1). However, a range of conditions exists (see the conceptual model discussion in Woods, 2015a). For example, river conditions may be gaining, throughflow, or losing when the river has low flow. The floodplain may or may not include a river lock (including a change in weir pool level due to the construction of the lock). The floodplain may be narrow or wide, and the dominant direction of regional groundwater flux may be lateral or vertical into the floodplain. For this reason, the Goyder Floodplain Model has six bases cases representing a variety of conditions encountered in the study area. This is explained in greater detail below.

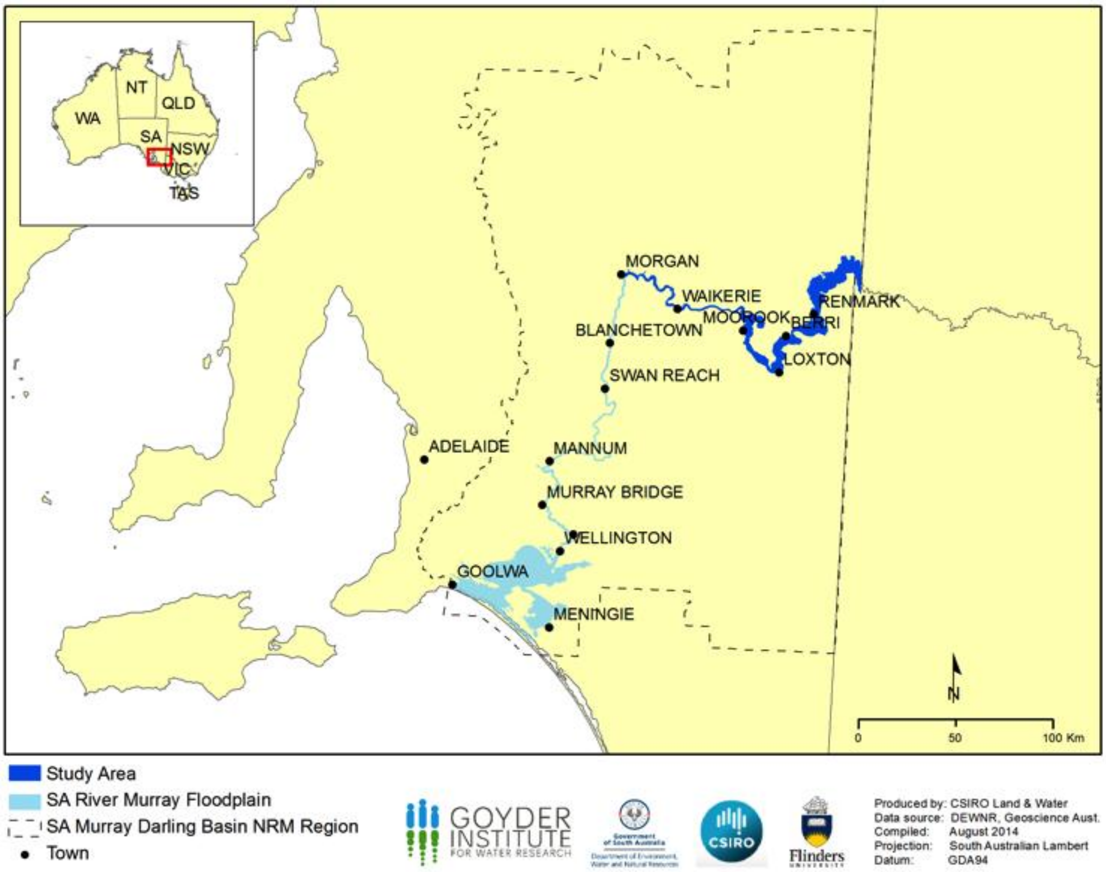


Figure 4-1 Model study area

Hydrogeological parameters vary within the study area. The base case parameters are informed by an amalgamation of data from this area, but are not specific to any one location. Where there is a wide range of possible parameters, values from the Pike Floodplain have been adopted. Pike Floodplain was selected as it is currently DEWNR’s highest priority for a numerical groundwater model, owing to the proposed regulator and salinity management works.

Where there is uncertainty regarding how to simulate a process or driver, the base case and initial scenarios adopt methods used by the Salinity Register models. The Salinity Register models are currently the most up-to-date simulations of SA MDB groundwater processes but simulate floodplain processes extremely simply, hence they provide a starting set of assumptions. It is anticipated that the Salinity Register models will be revised over time to include improved representation of floodplain processes, based on this study and others.

The Salinity Register models do not simulate solute transport (Section 1.3.1). Instead, the flux of solute is determined by assigning a set concentration, based on a background salinity value, to the gained river volume. This approach is sufficient for Salinity Register purposes, to estimate long-term river salinity impacts from changes on the highland, and the models have provided a reasonable match to observed salinity trends in the River Murray. However, solute transport is required to capture short-term, localised salinity impacts on the floodplain, such as those driven by environmental watering, natural flooding, floodplain pumping or weir-pool manipulation. Solute transport is included in the Goyder Floodplain models to simulate variation both spatially and temporally of the groundwater salinity distribution on the floodplain.

4.1.1 Hydrogeology

As the Goyder Floodplain Model is a generic SA MDB floodplain model, the hydrogeology represented in the model is also generic. At many locations in the study area, the hydrogeology is significantly more varied and complicated.

The hydrogeology of the floodplains in the lower River Murray is described in detail in Woods (2015a). Within the incised river trench, sediments consist of the Monoman Formation (a semi-confined aquifer) overlain by the Coonambidgal Formation (an aquitard). Beneath the Monoman Formation are the regional units: an aquitard such as the Bookpurnong Formation and the regional aquifer system of the Murray Group (Figure 4-2). The edge of floodplain (defined by the incised river trench) sits adjacent to a regional aquifer, either the Loxton Sands Formation or the upper units of the Murray Group, depending on location. For the sake of simplicity, the regional aquifer laterally adjacent to the floodplain aquifer is referred to as the Loxton Sands in this chapter; however, the model results are likely to be similar for locations where this aquifer is the upper units of the Murray Group.

The Coonambidgal Formation consists predominately of clay sediments and is known to be of low hydraulic conductivity at many locations along the lower River Murray. However, it is also fairly variable along the river channel, ranging from low-permeability clays to sandy clays and clayey silts. It also has a varying thickness, from absent to up to 10 m thick. The Coonambidgal Formation determines the unconfined to semi-confined nature of the Monoman Formation.

In terms of floodplain processes, the Coonambidgal Formation is critical in determining how groundwater will interact with surface processes, such as evapotranspiration, as well as how floodplain inundation becomes groundwater recharge. It is also in direct connection with the river for much of the floodplain as well as other surface water features.

The Monoman Formation is considered to be a semi-confined to unconfined aquifer consisting of clean fine to coarse alluvial sands (Anon, 1989). The thickness of this aquifer also varies across the floodplain, in the order of 10 to 40 m. In some areas it is known to be in direct connection with the river. The salinity of this aquifer ranges from 5,000 to 70,000 mg/l (RPS Aquaterra, 2012).

The Goyder Floodplain Model has two topographic conceptualisations. The first conceptualisation focuses on the incised river trench only (Figure 4-3 (a)) and includes a broad floodplain on either side of the river. The second conceptualisation includes a highland area on one side of the river, with the river close to the edge of the highland (Figure 4-3 (b)).

The Loxton Sands is a regional aquifer that sits on the highland adjacent to the Monoman Formation and Coonambidgal Formation. It is also a semi-confined to unconfined aquifer consisting of fine to medium sands with some clay and silt layers. The salinity of the Loxton Sands is similar to the Monoman Formation.

The regional aquitard, where present, separates the overlying Monoman/Loxton Sands and the underlying Murray Group. For example, the Bookpurnong Formation consists of poorly consolidated silts and clays and can be up to 40 m thick.

The Murray Group is a regionally extensive confined aquifer of varying water quality. It consists of multiple different formations, which vary in presence/absence throughout the floodplain. These formations also vary in hydraulic properties, with some formations being better aquifers than others. This group can be in the order of 100 m thick, depending on location and has a salinity of approximately 20,000 mg/l (Sharley & Huggan, 1995).

As discussed in Woods (2015a), both floodplains and surface water features can be gaining, throughflow or losing. The majority of floodplains in the study area would be considered gaining floodplains, where groundwater flows into them.

Table 4-1 Hydraulic parameters from previous studiesprovides a summary of hydraulic parameters for these units from previous studies and has been used to inform the parameters in the Goyder Floodplain Model.

Table 4-1 Hydraulic parameters from previous studies

<table>
<tr><th colspan="2" rowspan="2">Hydrogeological Unit</th><th colspan="2">Hydraulic conductivity (m/day)</th><th>Tranmissivity (m^2/day)</th><th>Storage (-)</th><th></th><th rowspan="2">Source</th></tr>
<tr><th>K_h</th><th>K_v</th><th>T</th><th>$S^{(1)}$</th><th>S_y</th></tr>
<tr><td colspan="2">Woorinen Formation</td><td>-</td><td>-</td><td>-</td><td>-</td><td>-</td><td></td></tr>
<tr><td colspan="2">Coonambidgal Formation</td><td>-</td><td>-</td><td>-</td><td>-</td><td>-</td><td></td></tr>
<tr><td colspan="2" rowspan="3">Monoman Formation</td><td>2 – 19</td><td>-</td><td>-</td><td>$4x10^{-4}$</td><td>-</td><td>URS (2000)
REM (2002b)
REM (2002a)</td></tr>
<tr><td>25 - 33</td><td>-</td><td>500 – 655</td><td>$5x10^{-3}$</td><td>-</td><td>AWE (2012a)
AWE (2012b)</td></tr>
<tr><td>4.07 – 536</td><td></td><td>30 – 1800</td><td>$7.06x10^{-7}$ – $6.47x10^{-2}$</td><td></td><td>AWE(2011)</td></tr>
<tr><td colspan="2">Blanchetown Clay</td><td>-</td><td>$3x10^{-4}$</td><td>-</td><td>-</td><td>-</td><td>REM (2002b)</td></tr>
<tr><td colspan="2" rowspan="3">Loxton Sand</td><td>0.3 – 5.8</td><td>-</td><td>26 – 121</td><td>$2x10^{-4}$ – $2x10^{-3}$</td><td>0.06-0.13</td><td>REM (2002b)
REM (2002a)
REM (2005)</td></tr>
<tr><td>3 – 59</td><td>-</td><td>79 – 1470</td><td>$5x10^{-6}$ – 0.275</td><td></td><td>AWE (2012a)
AWE (2012b)</td></tr>
<tr><td>2.59 – 33.4</td><td></td><td>41.5 – 421</td><td>$2.45x10^{-8}$ – $2.96x10^{-1}$</td><td></td><td>AWE(2011)</td></tr>
<tr><td colspan="2">Lower Loxton Clay and Shells</td><td>-</td><td>-</td><td>-</td><td>-</td><td>-</td><td></td></tr>
<tr><td colspan="2">Bookpurnong Formation</td><td>-</td><td>$6x10^{-4}$ – 1.6</td><td>-</td><td>-</td><td>-</td><td>REM (2002b)
REM (2002a)</td></tr>
<tr><td rowspan="6">Murray Group Limestone</td><td rowspan="2">Pata Formation</td><td>0.05 – 0.1</td><td>-</td><td>-</td><td>$1x10^{-3}$ (S_s)</td><td>-</td><td>REM (2002b)</td></tr>
<tr><td>0.09 – 0.72</td><td>$1x10^{-3}$ – $6x10^{-3}$</td><td></td><td>$3.12x10^{-5}$ – $4.91x10^{-4}$</td><td></td><td>AWE(2011)</td></tr>
<tr><td>Winnambool Formation</td><td>-</td><td>-</td><td>-</td><td>-</td><td>-</td><td></td></tr>
<tr><td>Glenforslan Formation</td><td>0.14 – 0.56</td><td>$3x10^{-3}$ – $10x10^{-3}$</td><td>-</td><td>$1.56x10^{-4}$ – $1.95x10^{-4}$</td><td>-</td><td>AWE (2011)</td></tr>
<tr><td rowspan="2">Mannum Formation</td><td>0.05 – 0.1</td><td>-</td><td>-</td><td>-</td><td>-</td><td>URS (2000)</td></tr>
<tr><td>1.25 – 2.56[(2)]</td><td>$62x10^{-3}$ – $133x10^{-3}$[(2)]</td><td></td><td>$1.48x10^{-4}$ – $2.75x10^{-4}$ [(2)]</td><td></td><td>AWE (2011)</td></tr>
</table>

(1) S is storage coefficient, which is the product of specific storage (1/m) and layer thickness (m), and is hence dimensionless

(2) Upper Mannum Formation

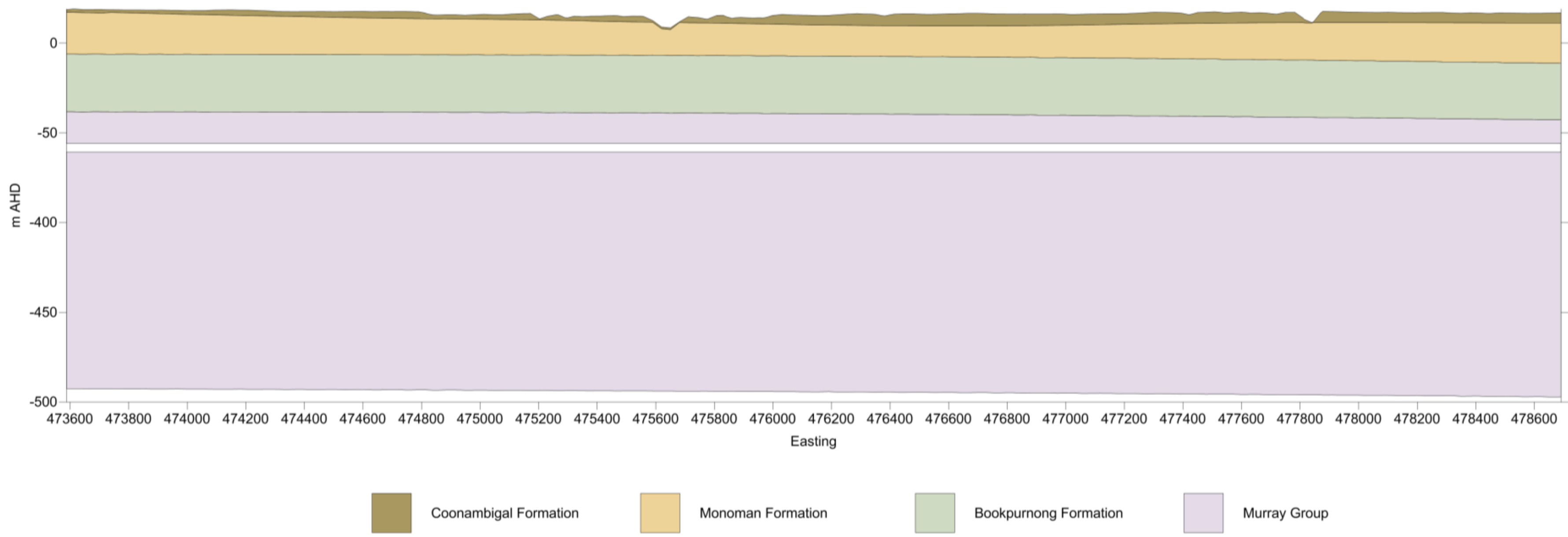


Figure 4-2 Simplified Floodplain Hydrogeology

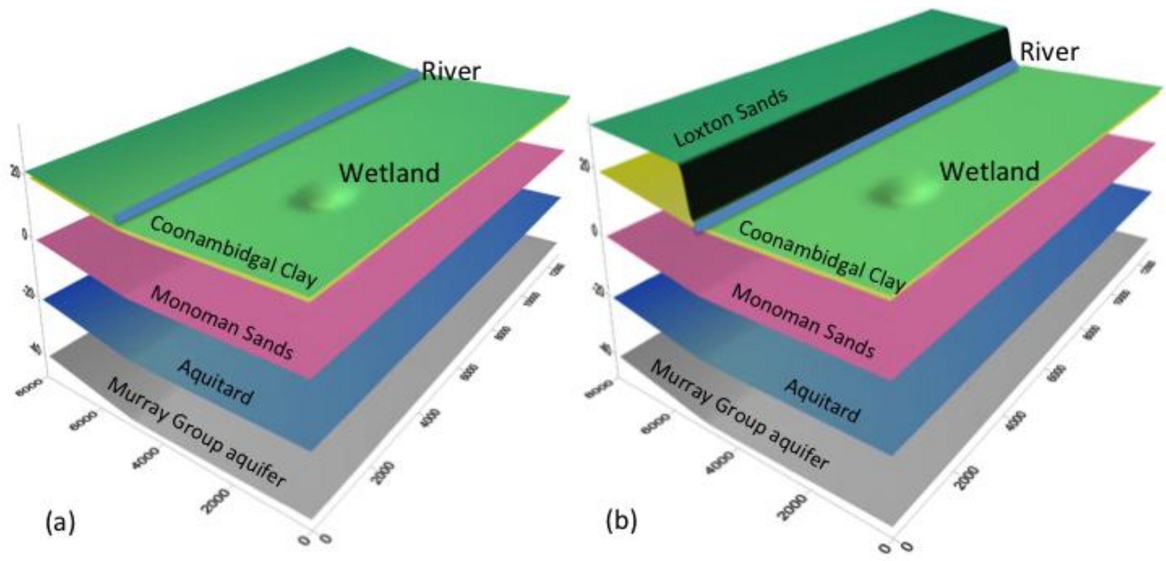


Figure 4-3 Goyder Floodplain Model Conceptualisation ((a): Broad Floodplain and (b): Broad Floodplain plus Highland)

4.1.2 Surface Water

Surface water plays a key component in how floodplains behave and interact with the groundwater. Surface water features are simulated simply within the Goyder Floodplain Model, consisting of the main river channel and a wetland. The River Murray in the study area may have gaining, throughflow or losing conditions, hence it was necessary to develop base cases which covered all these possibilities.

4.1.3 Base case conceptualisations

From the discussion above, it was identified that a minimum of three model conceptualisations needed to be represented in the Goyder Floodplain Model. In addition to the three conceptualisations, the river needed to be represented as both a river with a lock and a river without a lock. From these combinations, the base models given in Table 4-2 were developed.

Table 4-2 Goyder Floodplain Model – Base Model Conceptualisations

Case	Floodplain	River	Lock present
Case A – with Lock	Gaining	Losing	Yes
Case A – without Lock	Gaining	Losing	No
Case B – with Lock	Gaining	Throughflow	Yes
Case B – without Lock	Gaining	Throughflow	No
Case C – with Lock	Gaining	Gaining	Yes
Case C – without Lock	Gaining	Gaining	No

4.2 Base Model Construction, Validation and Initial Findings

4.2.1 Model code

The Goyder Floodplain model has been built in Groundwater Vistas Version 6. Initially, MODFLOW2000 (Harbaugh *et al.* 2000) was used to simulate groundwater flow, but this led to difficulties with re-wetting that were resolved by switching to the MODFLOW2005 (Harbaugh, 2005) code using the NWT solver and UPW package (Niswonger *et al.*, 2005) (Section 4.2.8.2). MODFLOW-USG (Panday *et al.*, 2013) was used to investigate spatial discretisation of the model domain (Section 4.3.1.3). MODFLOW Surfact (Hydrogeologic, 1996) was used to investigate re-wetting of the Coonambidgal Formation (Section 4.2.8.2). The NWT automate solver options was set to moderate and the solver used a head change criteria of 10^{-5} m, a flux change criteria of 10 m^3/day and maximum outer iterations set to 100.

The MODFLOW-related solute transport code MT3DMS (Zheng & Wang, 1999) was used to simulate solute transport for almost all flow scenarios. MT3DMS utilizes the flow fields generated during the MODFLOW simulation to solve the advection-dispersion equation. MT3DMS utilizes an implicit formulation based on generalized conjugate gradient (GCG) method to solve the dispersion and sink/source terms and a separate solution scheme for the advection term selected by the user. All simulations utilized the inherently stable third-order mass-conservative TVD (total variation diminishing) scheme to solve the advection term with the Courant number criteria restricted to 0.75. The concentration closure criterion for convergence with the GCG solver was set at 10^{-9}. Transport step sizes remain fixed for each simulation and are determined internally by the TVD scheme, which avoids any potential solution oscillation during iteration as well as minimal numerical dispersion. Sorption was excluded from all solute transport simulations.

4.2.2 Domain and Grid

The model domain for the Goyder Floodplain model is loosely based on the dimensions of the Pike Floodplain and is 13 km long by 8 km wide. The cell size for the model is 125 m x 125 m, which was chosen based on the cell sizes used in other models that cover the lower River Murray floodplain (notably the SA Salinity Register Models). This resulted in a model with 64 rows and 104 columns with a total of 26,624 active cells over 4 layers.

The impact of cell size was investigated and is reported in Section 4.3.1.3.

4.2.3 Initial conditions and stress periods

Initial conditions for the base models set potentiometric head at top of surface, then ran until convergence at steady state. The heads from the steady state model were then used as initial conditions for the scenario modelling. Where the scenario changed a key parameter, such as evapotranspiration extinction depth, the base models were re-run with this parameter change to create a new initial head for that scenario run.

Initial solute distributions for scenario models were obtained by running MT3DMS to a perceived steady state using the steady state flow field from the relevant MODFLOW base simulation. A difference of less than 0.001% in the salt flux to river for consecutive months was selected as the metric for perceived steady state. Evapoconcentration was included in all simulations and resulting in greater concentrations of 35,000 mg/l on the floodplain. To facilitate better comparison with the current method used to determine salt flux to river all concentrations exceeding 25,000 mg/l in the Monoman Formation were reduced to 25,000 mg/l for the initial solute distributions.

All the transient scenarios were run over a 20 year time period, nominally based on the period between 1/1/1992 and 31/12/2011. This time period was selected as it includes both a wet period with significant flooding as well as an extended dry period. Where appropriate, data from this period has been incorporated into the model. Stress period length depends on the scenario. Annual, monthly, or sub-monthly stress periods were used.

4.2.4 Model Layers

The Goyder Floodplain Model has four layers, in line with the hydrogeological conceptualisation outlined previously (Figure 4-3). Layers 2 to 4 are consistent between model base cases, while layer 1 is consistent between Cases A and B but modified in Case C.

4.2.4.1 Surface Elevation

The Goyder Floodplain Model, while being designed to represent floodplain conditions, is also a synthetic model, developed to investigate scenarios as efficiently as possible given the time constraints of the project. These constraints played a significant role in the synthetic surface elevation developed for the model. The reason for a synthetic surface elevation was to remove the complicated interaction topography has with floodplain processes. For example, topography controls surface water flow, evapotranspiration and hence the watertable. By creating a simplified surface elevation for the model, it was intended that the differences due to processing or modelling implementation would be easier to interpret.

The synthetic elevation data was loosely based on surface elevations at Pike Floodplain, however it was also constrained by the need to implement a lock, where the river height varied by approximately 4 m, preferably without the river sitting above the surface of the model. The river was implemented in layer 2; however, as the lower layers followed the same gradients as the top layer (e.g. the elevation of the top of layer 2 was the surface elevation minus 2 m) any gradient in the surface elevation needed to account for this variation in river level.

Surface elevation of the Goyder Floodplain Model has therefore been constructed to have a right to left gradient of approximately 2%. As shown in Figure 4-4 and Figure 4-5, the elevation in the centre of the model drops from 18 m AHD on the right to 15 m AHD on the left. This gradient allowed for some degree of in-channel river fluctuation without the river sitting unrealistically above the surface of the model.

The surface elevation has been constructed with the lowest elevation at the horizontal mid-line of the model, which is flat either side of this mid-line for about 500 m. This area was nominal location of the river and provided the flexibility to move the river within this zone while still allowing the river to be in the lowest point in the landscape. From this low point, the surface elevation rises by 6 m to the edge of the model on the north of the river (not actual north, see Figure 4-4 for nominal direction indicators), and 1 m to the south of the river for Case A and B (Figure 4-4). Case C built on this configuration, however the area north of the river steeply rose, representing cliffs, to a height of 18 m above the lowest surface elevation (Figure 4-5).

One other addition was made to the surface elevation to create a wetland area, south of the river. This was designed to fill and store water during and after inundation events in a simpler version of real world processes.

4.2.4.1 Layer Structure

Layer 1 is representative of the Coonambidgal Formation in Cases A and B and a combination of the Coonambidgal Formation on the floodplain and the Loxton Sands Formation on the highland in Case C. For Cases A and B as well as the floodplain area in Case C, layer 1 (Coonambidgal formation) is 2m thick. The thickness was developed by looking at elevation data of the surface and top of Monoman Formation from the Atlases (AWE, 2009; AWE, 2011; AWE, 2012a; AWE, 2012b; AWE, 2013). This data showed a range of thicknesses (0 to 10 m) throughout different floodplains and within floodplains. Pike Floodplain was looked at in more detail and thickness plot was developed for the Coonambidgal Formation in this area (Figure 4-6). This plot showed that a 2 m thickness for the Coonambidgal Formation was a reasonable assumption.

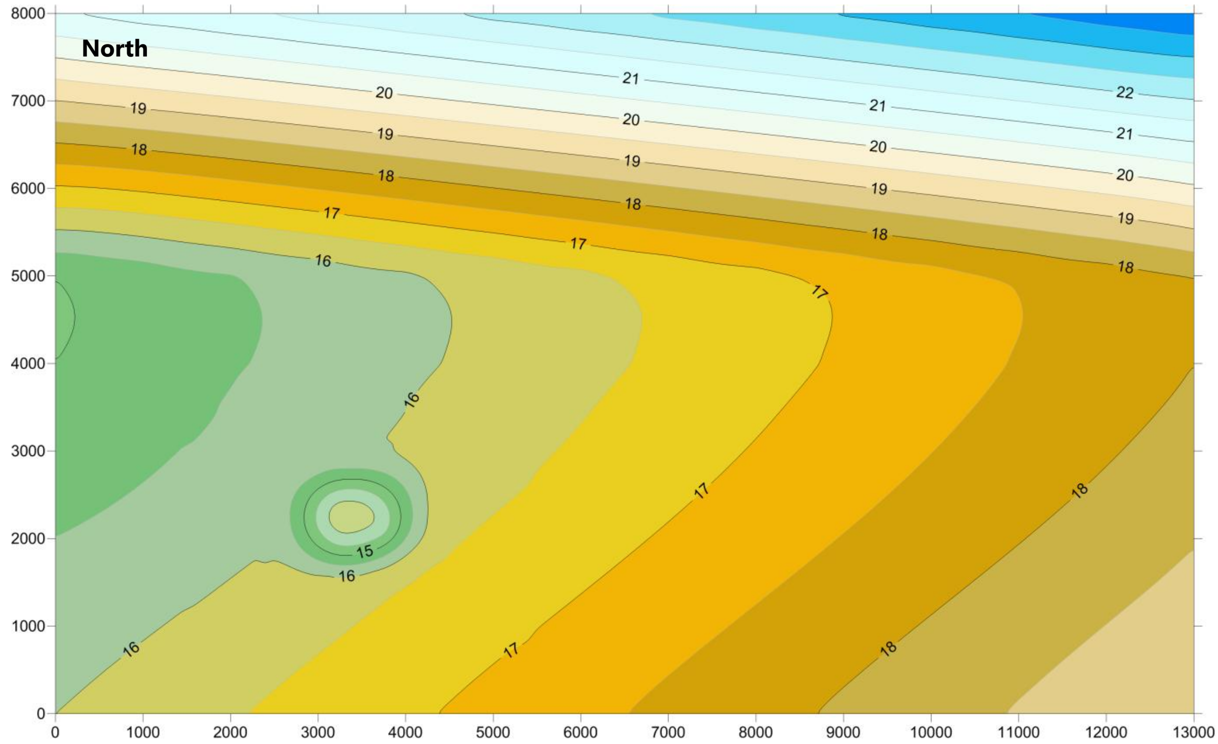


Figure 4-4 Surface Elevation Contours Case A and B

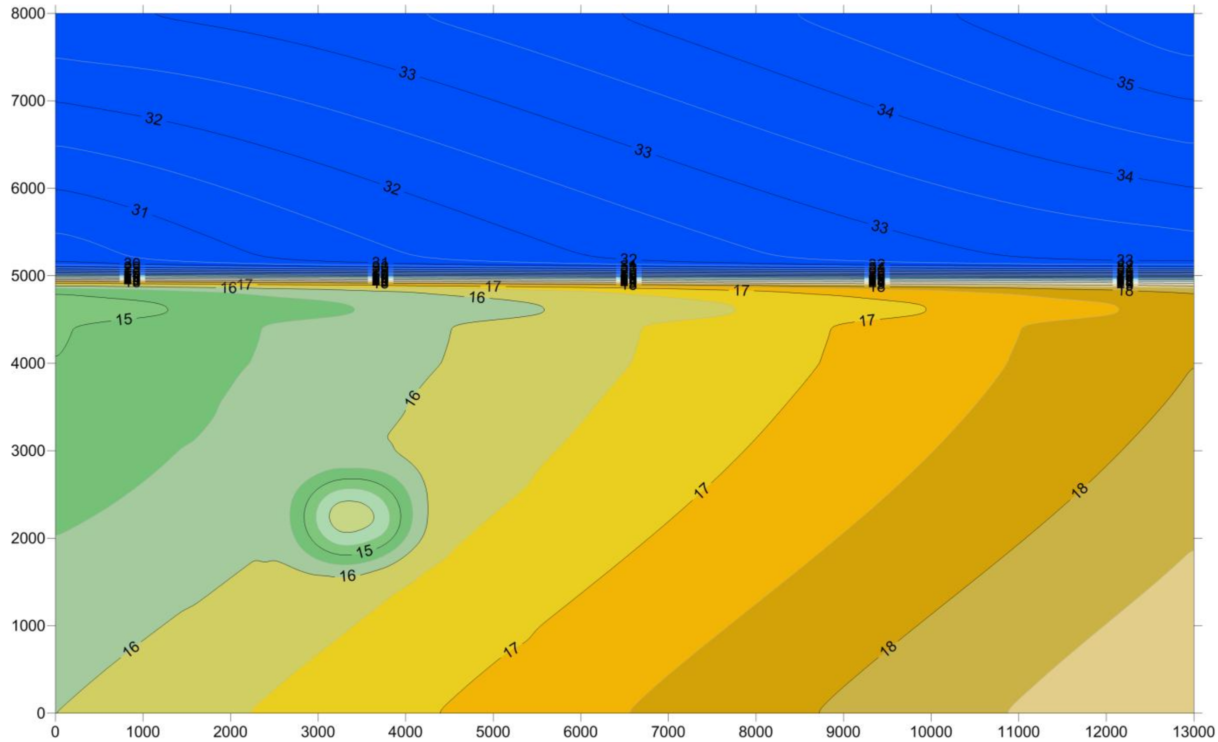


Figure 4-5 Surface Elevation Contours Case C

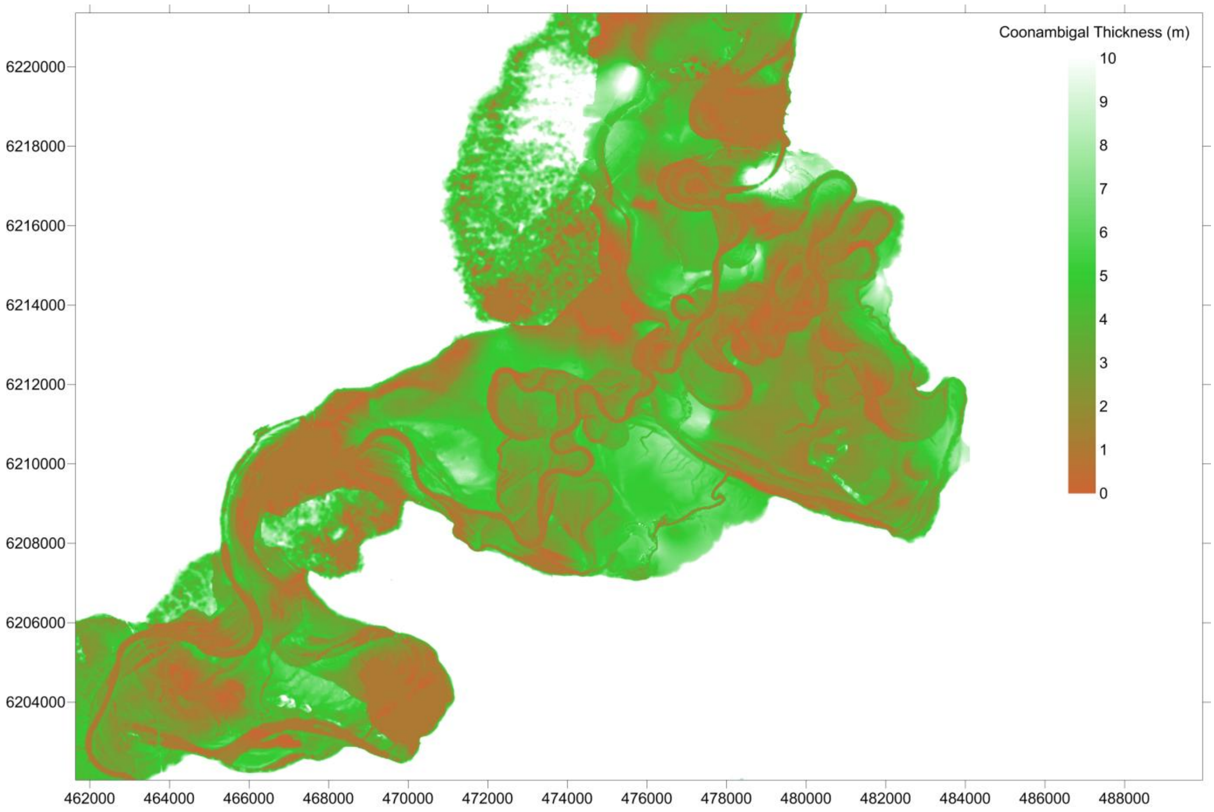


Figure 4-6 Thickness of Coonambidgal at Pike Floodplain based on existing bore data

For Case C, the addition of a highland area meant that a section of the model in layer 1 was required to have an increased thickness. This was based on observed differences in elevation between the floodplain and the highland area in the Atlases. From this information, an increased thickness of 20 m was used in the highland area of Case C.

Layers 2 to 4 represent the Monoman Formation, regional aquitard, and the Murray Group respectively. Each of these layers are 20 m thick, which is an approximation based on the data shown in the Atlases. It is acknowledged that for some floodplains in the study area, these units may vary significantly in thickness. In some cases, the regional aquitard may be completely absent. However, the Goyder Floodplain Model would require only small modifications to simulate these locales.

4.2.5 Hydraulic Parameters

Groundwater flow model parameters are summarised in Table 4-3. Most are based on aquifer test results for the Pike-Murtho region, as summarised in Woods *et al.* (2014). They are also broadly consistent with regional parameters values adopted in the Salinity Register models of the Border to Lock 3 (e.g. Woods *et al.,* 2014). The Coonambidgal Formation hydraulic conductivities are on the lower side of the estimates discussed in Woods (2015a).
Little data was available for the following parameters:

- Coonambidgal Formation storage and specific yield
- Monoman and Loxton Sands vertical hydraulic conductivity
- Regional aquitard horizontal hydraulic conductivity
- Regional aquitard storage and specific yield
- Murray Group vertical hydraulic conductivity
- Murray Group specific yield

Table 4-3 Hydraulic Parameters used in the Goyder Floodplain Model

Layer	Unit	K_x/K_y (m/day)	K_z (m/day)	S*	S_y
1	Coonambidgal	0.05	0.005	0.01	0.02
1	Loxton Sands (Case C only)	3	0.1	0.0001	0.15
2	Monoman Formation	20	0.2	0.0001	0.15
3	Regional Aquitard	0.006	0.0006	0.0001	0.15
4	Regional Aquifer (Murray Group)	0.5	0.05	0.0001	0.15

*S is Storage coefficient, which is dimensionless

In terms of units where either the horizontal or vertical conductivity was unknown, an assumption was made that the ration between horizontal and vertical hydraulic conductivity was 10:1 respectively. The only exception to this was the Monoman Formation aquifer where a ratio of 100:1 was used. This was primarily due to a vertical hydraulic conductivity of 2 m/day being unusually high, in addition to preventing excessive contrast in vertical hydraulic conductivity between layers. Because of this the Loxton Sands vertical hydraulic conductivity is slightly less than the 10:1 ratio would have calculated, since this formation is thought to have a slightly lower hydraulic conductivities than the Monoman Formation.

There are two other variations in the hydraulic conductivity that are not based on physical parameters but are employed to reduce model instability. The first variation is a zone of Monoman Sand parameters in layer 1 that sit directly above the cells hosting the river in layer 2 (Section 4.2.6.2). The second variation is only applicable to Case C and is a hybrid zone in layer 1, two cells wide, between the Coonambidgal Formation and the Loxton Sands. The horizontal and vertical hydraulic parameters in this zone are 0.1 and 0.01 m/day respectively and represent a hybrid of the Coonambidgal Formation and the Loxton Sands parameters. This was used to prevent model instability that can occur when parameters that differ by orders of magnitude about each other.

The adopted parameters represent a locale where the predominant direction of regional groundwater into the floodplain is lateral. This is true for the study area east of Hamley Fault (i.e. upstream of Lock 3). To simulate locales where the predominant inflow direction of the regional flux was upwards, such as Woolpunda, the vertical hydraulic conductivity of the regional aquitard should be decreased. The regional boundary conditions would also need to be altered.

Parameters used in the solute simulations are listed in Table 4-4. Diffusion and dispersion parameters are applied to all layers and hydrostratigraphic units. The value entered for salt diffusion is typical for units of meters and days (Zheng & Wang, 1999). Dispersivity values are considered suitable for the scale of the problem according to (Gelhar *et al.*, 1992; Schulze-Makuch, 2005). Values of effective porosity are elevated slightly above field-derived specific yield values and were found to provide a stable solution with adequate time step size. A reduction in effective porosity by 0.1 for all geological resulted in unacceptable levels of numerical dispersion evidenced by excessively large salt concentrations at convergent flow locations.

Table 4-4 Solute transport model parameters

Parameter	Value	Unit
Diffusion	8.64e-05	m^2/day
Longitudinal Dispersion	65	m
Transverse Dispersion	20	m
Longitudinal Dispersion vertical	6.5	m
Transverse Dispersion vertical	20	m
Effective porosity Coonambidgal	0.35	-
Effective porosity Monoman	0.3	-
Effective porosity aquitard	0.35	-
Effective porosity Murray Group	0.3	-

4.2.6 Boundary Conditions

4.2.6.1 Regional groundwater flow

General head boundaries were used to represent the driving heads at the edge of the floodplain in the Monoman Formation Aquifer and Murray Group. As such, they were implemented in layers 2 and 4 on the northern and southern edges of the model (Figure 4-7). Case C has an additional general head boundary in layer 1 representing the groundwater head in the Loxton Sands at some distance from the floodplain.

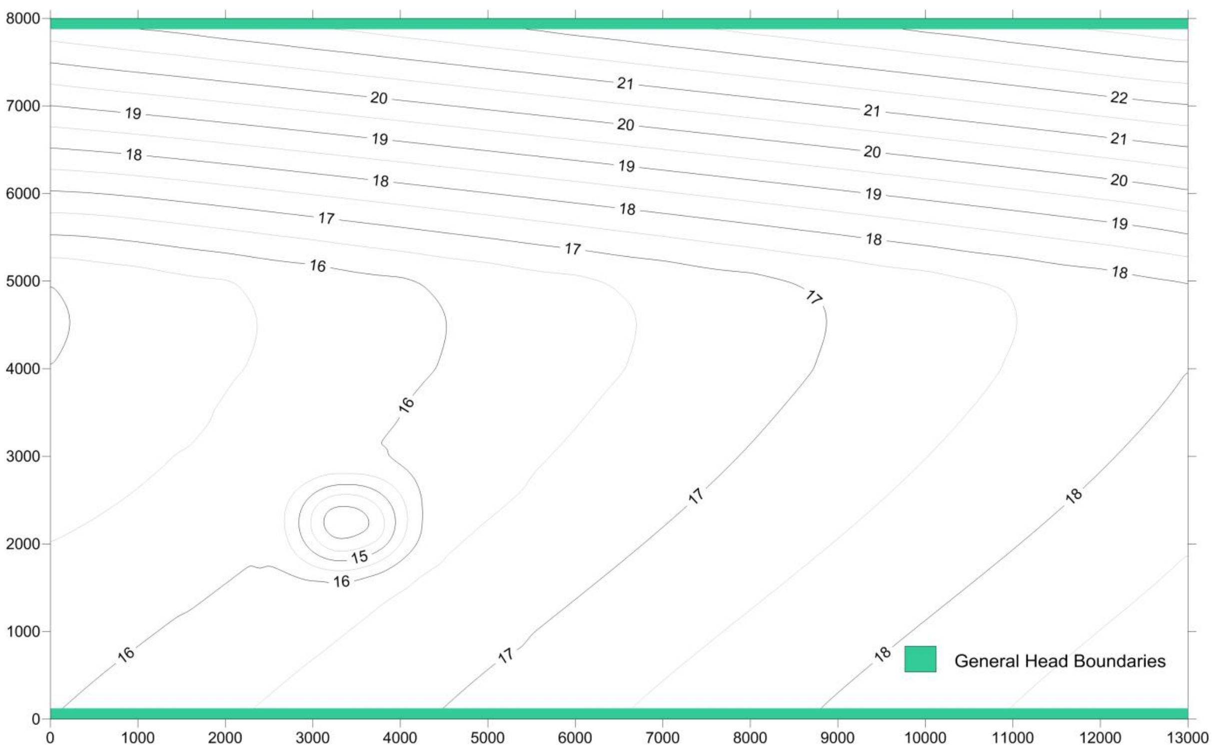


Figure 4-7 General Head Boundary Locations

The general head boundaries, in addition to the river location, were used to drive the representation of the desired conceptual system. Hence each case has a different configuration of heads, given in Table 4-5.

Table 4-5 General Head Boundaries

GHB Boundary Location	Case A	Case B	Case C
L1 North Head	N/A	N/A	L2 elevation* +7m (13.5 m above River stage**)
L2 North Head	L2 elevation* - 2m (4m above river stage**)	L2 elevation* - 1m (5m above river stage*)	L2 elevation +7m (13.5 m above River stage**)
L2 South Head	L2 elevation*	L2 elevation*	L2 elevation** + 1m
L4 North Head	surface elevation* + 4m	surface elevation* + 4m	Same height as Cases A and B (note topography is different in the north for this model).
L4 South Head	surface elevation* + 4m	surface elevation* + 4m	surface elevation* + 4m

*Note: Elevation refers to the elevation at the location of the boundary cell, which is either the northern or southern edge of the model.

**Note: River stage here refers to the stage in the models without a Lock

For the Monoman Formation (layer 2) a range of groundwater head boundary values were estimated from the Atlases and chosen based on the ability to represent the system conceptualisation. Due to the use of the synthetic topography, the head values implemented in the model are related to the relationship between the head and the river level observed in the Atlases, rather than actual measured head values. This was due to the head difference between the groundwater and the river being identified as critical for changes in groundwater system, rather than the difference between topography (at the edge or outside the floodplain) and the groundwater.

The values used in the model are consistent with differences in measured values between the highland heads and the river levels for the areas that each of the models are (generically) representing. Generally, for the losing and throughflow floodplains, the difference between the regional watertable and river level is 2 to 5 m, whereas for a gaining river situation the difference is in the order of 12 to 15 m. These heads were further constrained by the model validation process outlined in Section 4.2.7.

For the Murray Group, general head boundary values were based on observed heads in the Murray Group compared to the elevation of the floodplain. For the Pike floodplain on the southeastern side, the difference between the elevation across the floodplain edge and the potentiometric head ranges between 3 and 8 m, based on spot heights of the elevation and the watertable in the Murray Group (AWE, 2012b). The value chosen for the model is consistent with these observations.

Conductances for boundary conditions are generally not a measured property, but are inferred from hydraulic conductivities. The conductances used in the Goyder Floodplain Model are 250 m^2/day and 60 m^2/day for layer 2 (and layer 1 in Case C) and layer 4 respectively.

This conductance for layer 2 (and layer 1 in Case C) is based on a horizontal hydraulic conductivity of 10 m^2/day, which is a hybrid horizontal hydraulic conductivity representative of the average between the Monoman Formation and the Loxton Sands Formation. The conductance for layer 4 was calculated based on a horizontal hydraulic conductivity of 0.5 m^2/day, which is consistent with the horizontal hydraulic conductivity used in the Murray group formation that this boundary sits in.

The modelled regional flux into the floodplain was found to be broadly consistent with values estimated by Salinity Register models for the Border to Lock 3 region. West of Lock 3, the predominant regional fluxes are likely to be upwards rather than lateral.

The salinity of regional groundwater entering the model is 25,000 mg/l in layer 2 and 20,000 mg/l in layer 4, as these are values typical for the eastern portion of the study area. The GHB boundary condition is of Cauchy type, that is, flow entering the domain enters with the designated concentration but flow exiting the domain has model-derived concentration. It should be noted that MT3DMS does not account for diffusion or dispersion across a Cauchy boundary.

4.2.6.2 River

Depending on location, the River Murray may be situated wholly in the Coonambidgal Formation, wholly in the Monoman Formation, or within both. Where the River Murray is wholly in the Coonambidgal Formation, the Coonambidgal Formation is generally thicker than in the other two cases (AWE, 2011). In addition, the further downstream, the more likely it is to reside in the Monoman Formation.

From this investigation, it was decided that hosting the river in the Monoman Formation was consistent with the stratigraphy and layer thickness used in the model. An additional consideration was given to the potential for convergence issues that may occur if a river is hosted in a formation of low hydraulic conductivity, such as the Coonambidgal Formation. Therefore the river boundary condition was implemented in layer 2 for all cases.

An issue with placing the river in layer 2 was the appropriate parameterisation of the cells above the river in layer 1. Ideally there would be no cells above the river; however, MODFLOW does not have this capability. Initially, parameters representative of air were used, but this caused numerical instability (this was prior to using MODFLOW2005 with the NWT solver and UPW package). Monoman Formation parameters were then adopted as they were considered high enough that they were consistent with the concept of having open space above the river, but low enough not to cause instability.

The river was constructed as 250 m, or two model cells, wide. This allowed for the inputs/outputs to the river to be unpacked for both the northern and southern sides of the river. It is also consistent with the approximate width of the River Murray in the study area.

The location of the river within the model domain varies between cases (Figure 4-8, Figure 4-9, Figure 4-10). In Case A, the river is in the centre of the model, which allowed the river to be losing as the wide floodplain of either side of the river lowered groundwater heads via evapotranspiration. Case B has the river further north, reducing the width of the floodplain to create throughflow river conditions. Case C has the river placed close to the cliffs to create a gaining river situation.

The depth of the river was based on NanoTem data (Telfer *et al.,* 2004), which show that the river is approximately 5 m deep. While there is variation in this, particularly at the locks, this is consistent with the approach taken in the Salinity Register models.

River level (river stage in MODFLOW), for the versions of the model which include a lock, was based on the upstream and downstream pool levels at Lock 5. Pool levels are used for steady-state models and initial conditions, i.e. 16.3 m AHD upstream and 13.2 m AHD downstream. For the versions of the model without a lock, river level was set at 2 m below surface elevation. This was based on comparing river level to spot heights nearest to the river in the Atlases, at locations away from the locks. River levels were held constant in the base case models but varied over time in some scenarios (Section 4.3.1).

The river was simulated using MODFLOW river cells in most cases. MODFLOW reservoir cells were investigated in some scenarios (Section 4.3.1.1).

No direct measurements have been made of riverbed conductance. A riverbed conductance of 1,000 m^2/day was used in the model. This is similar to the conductance used in the Salinity Register models for the main river channel, 1,500 m^2/day, but was lowered to minimise numerical instability (Section 4.2.8.3).

The salinity of the river water is 368 mg/l, based on an average given in (AWE, 2010). Similar to the GHB boundaries discussed in the previous section, river and reservoir boundary conditions are Cauchy-type and do not account for diffusion or dispersion across the boundary.

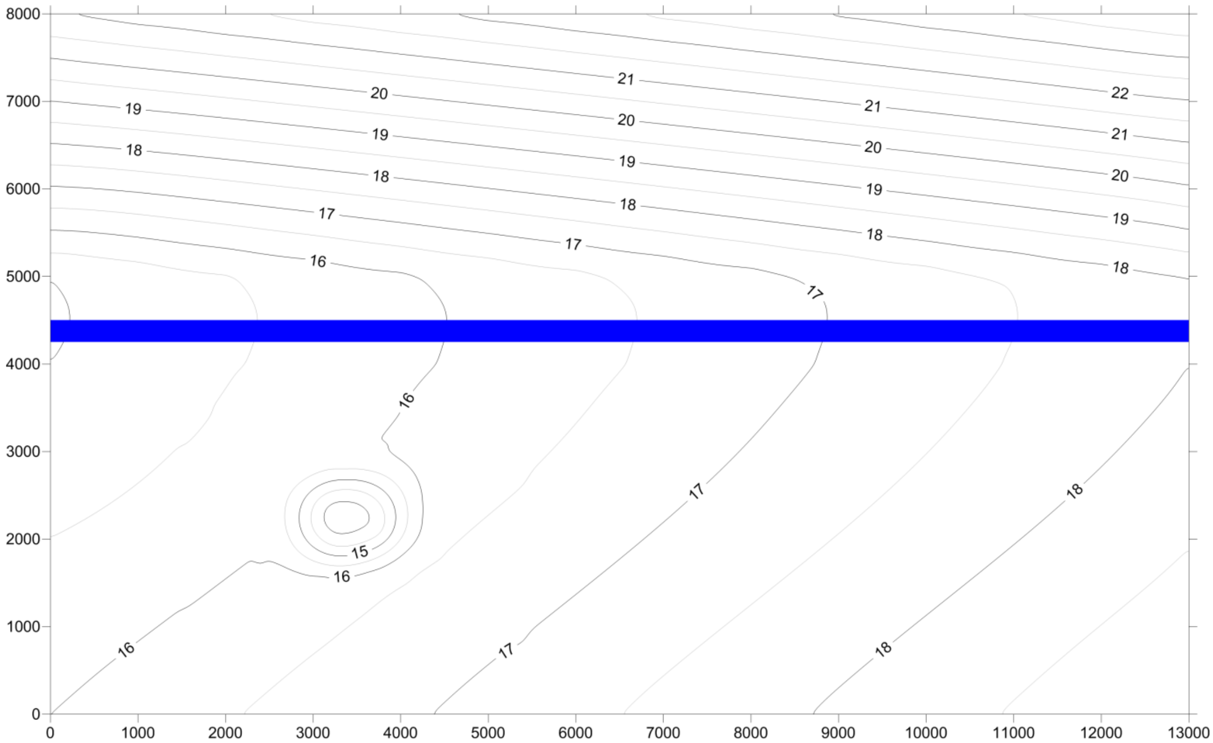


Figure 4-8 Case A River position

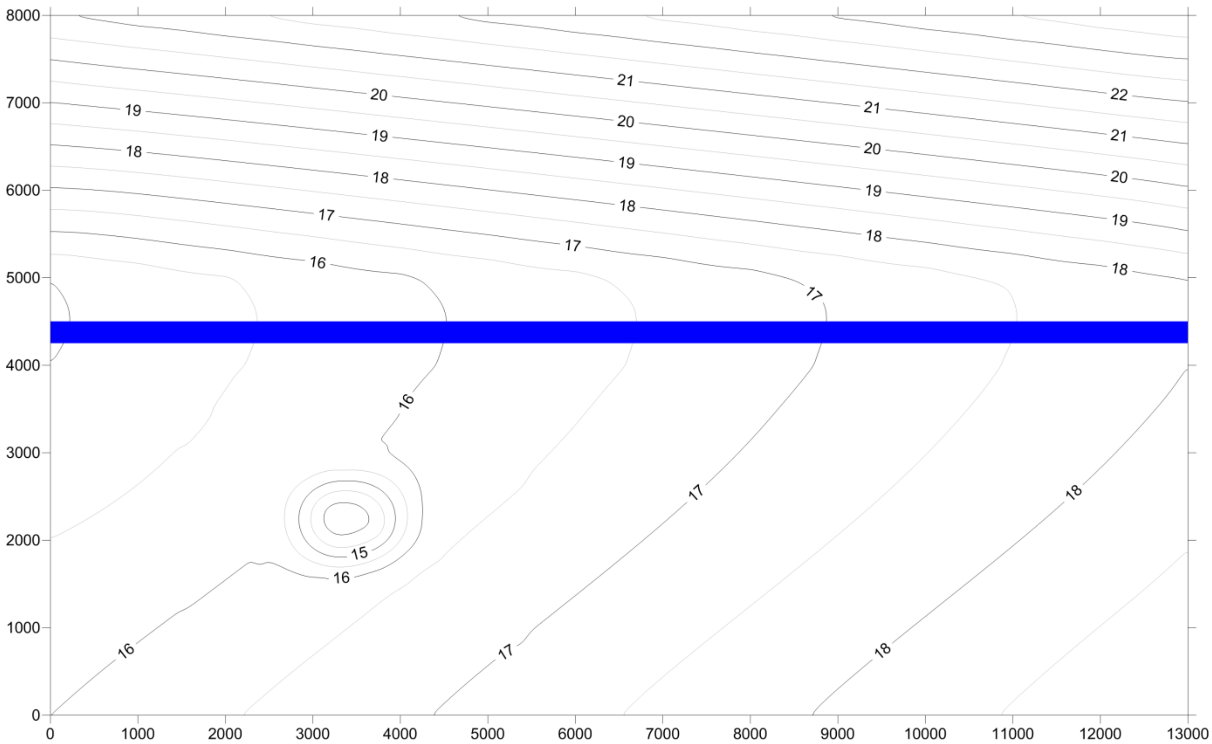


Figure 4-9 Case B River position

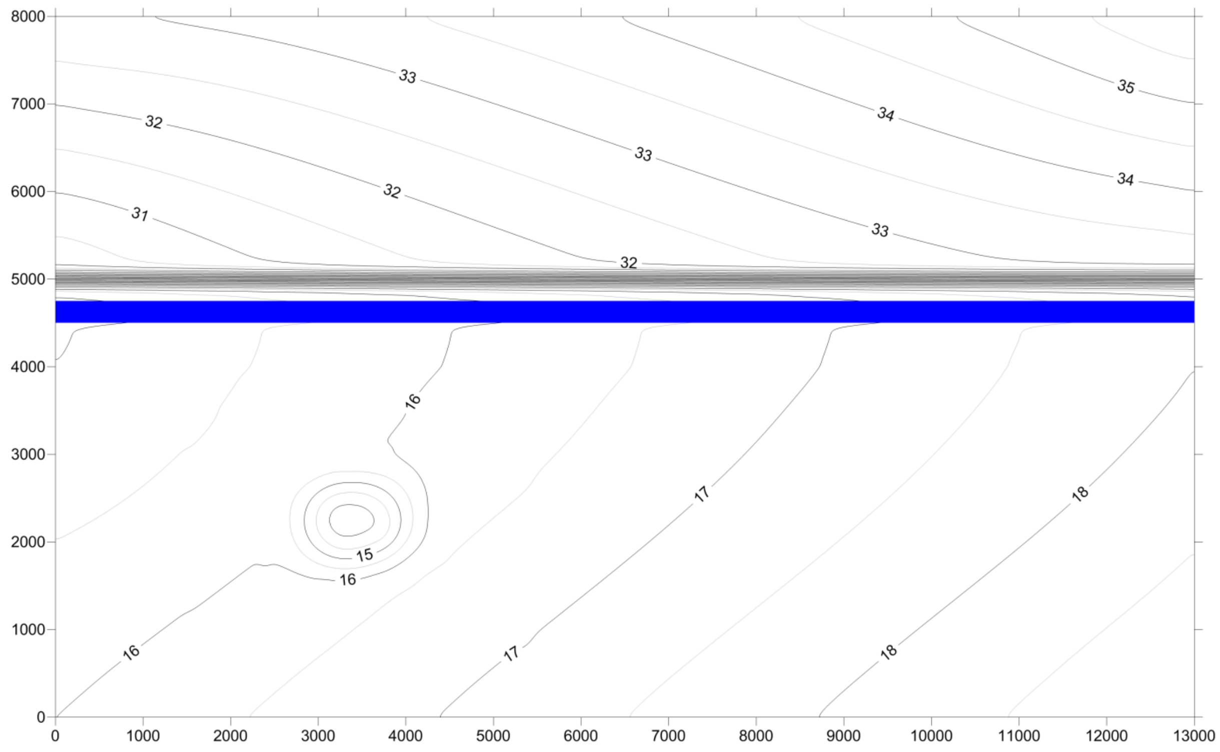


Figure 4-10 Case C River position

4.2.6.3 Recharge and Evapotranspiration

Recharge and ET are constant over time and do not vary spatially in the base case models. Recharge due to inundation is simulated in some scenarios (Section 4.4.3) while ET is varied in others (Section 4.4.2).

Groundwater recharge in the Goyder Floodplain Model is considered to be minimal (recharge due to inundation will be discussed in Section 4.3.3). As such, a background recharge of 0.1 mm/yr was used, based on estimates of mallee recharge (Allison *et al.*, 1990). This value is consistent with the background recharge used in the Salinity Register models. As an initial approximation, recharge is given a salinity of 0 mg/l, which is definitively an underestimate.

Potential evapotranspiration from groundwater was set at 1100 mm/yr with an extinction depth of 2.5 m. The standard linear MODFLOW function for ET was adopted for most simulations. This is consistent with the Salinity Register models, which simulate actual ET within observed ranges (e.g. see Woods *et al.*, 2014). It also allows ET to impact the Monoman Formation in Layer 2, given that the overlying Coonambidgal Formation layer is 2 m thick. There is a degree of variation in evapotranspiration extinction depth within the literature and floodplain models, dependent on vegetation type and other factors. To address this, sensitivity to ET parameters is considered in Section 4.3.2. The shape of the ET function with respect to depth to water is also investigated in Section 4.3.2.

4.2.7 Validation of Base Models

While the Goyder Floodplain Model is a synthetic model, in that it does not model a specific location, attempts were made to constrain the groundwater flow models based on the current knowledge of floodplains along the SA River Murray. The constraints on the system included:

- Representing gaining floodplains and river conditions which may be gaining, throughflow, or losing
- Actual evapotranspiration out of the model consistent with estimates of floodplain evapotranspiration (~60 to 80 mm/yr) (Holland *et al.*, 2011).
- Salt load to river less than 10 tonnes/day/km (based on Run of River surveys and Salinity Register model results)

In addition, as the hydraulic parameters are those of a locale where the majority of the regional flux comes from the laterally-connected aquifer:

- Inflow in general head boundaries in Monoman Formation (and Loxton in Case C) greater than inflow in general head boundaries in Murray Group

The base case models were run under steady-state conditions and a summary of the results based on these constraints is given in Table 4-6. In general the models are approximately consistent with the constraints outlined above. The one exception is the salt load to the river for Case C, which is slightly higher than the 10 tonnes/day/km identified as a constraint.

Table 4-6 Model constraint results for models without locks

Constraint	Case A	Case B	Case C
Representing Conceptualisation	Yes	Yes	Yes
Inflow from Monoman (and Loxton) (m^3/day)	84	105	146
Inflow from Murray Group (m^3/day)	36	35	17
Actual Evapotranspiration from Model (mm/yr)	57	54	79
Salt Load to River (tonnes/day/km)	0.02	2.7	10.9

These constraints were only applied to the models without locks, with the locks then added to the constrained models. The effect of the locks can be seen in Table 4-7. There is very little change in the flow from the general head boundaries, however both the evapotranspiration and the salt load to river increase significantly. The increase in evapotranspiration comes from the increase in water table upstream of the lock and is driven by the river level (Figure 4-11 and Figure 4-12).

Table 4-7 Model constraint results for models with locks

Constraint	Case A	Case B	Case C
Inflow from Monoman (and Loxton) (m^3/day)	83	104	144
Inflow from Murray Group (m^3/day)	36	35	17
Actual Evapotranspiration (mm/yr)	75	70	124
Salt Load to River (tonnes/day/km)	10.4	13.8	10.9

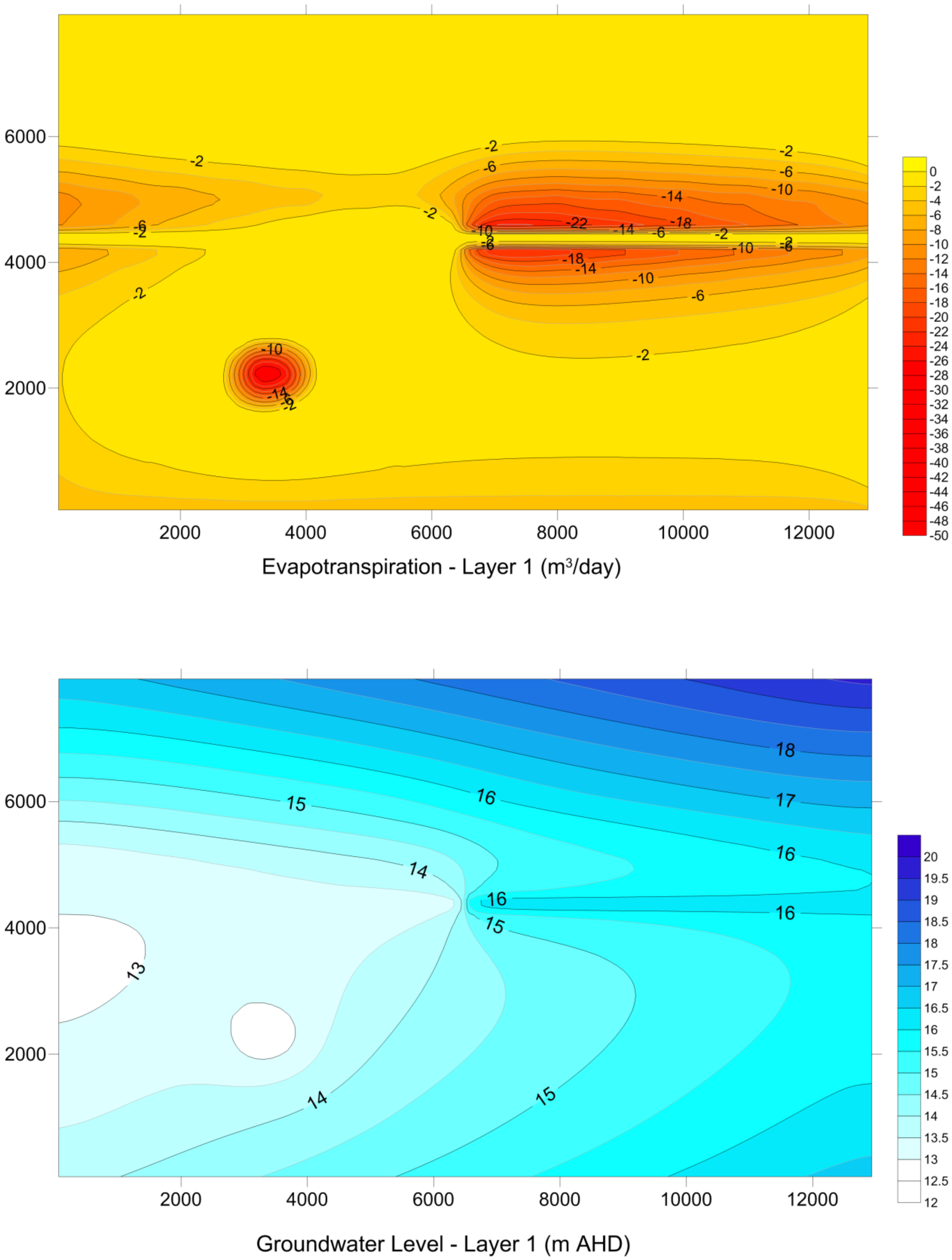


Figure 4-11 Actual evapotranspiration out and groundwater level for Case A with lock

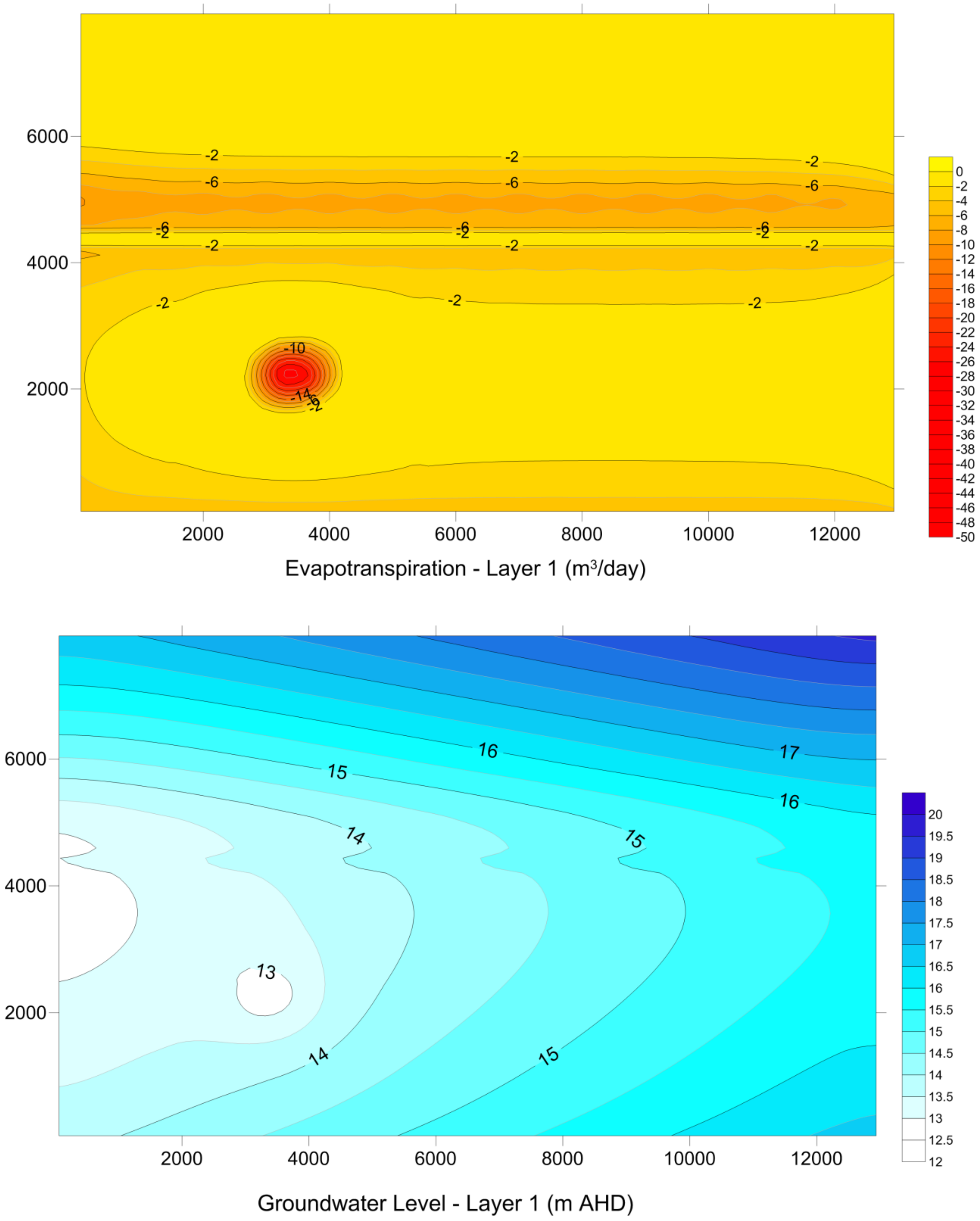


Figure 4-12 Actual evapotranspiration out and groundwater level for Case A without lock

Only the groundwater flow MODFLOW model was tested in this validation phase, so the salt load is calculated as flux to river multiplied by groundwater salinity of 20,000 mg/l. Interpreting this as an increased salt load may be misrepresenting the changes that occur due to the addition of the lock. Much of this increased flux is flow around the lock driven by the change in river water level, and is likely to be relatively fresh. For this reason, it is likely that the salt loads estimated in Table 4-7 are higher than what might be expected when simulating solute transport. This is supported by the results discussed in Section 4.4.

4.2.8 Findings from Base Models

4.2.8.1 Topography and river condition

Initially, all the base case models had a Case A topography, with the river distant from the floodplain edge within a wide floodplain. Under these conditions, and using the standard linear MODFLOW function for ET, it was extremely difficult to achieve gaining conditions for the River Murray. This was due to ET lowering potentiometric head in the floodplain aquifer. The location of gaining reaches was reviewed from the NanoTEM surveys (Telfer *et al.,* 2004) and it was found that they occurred only when the river was close to the edge of the floodplain, or where there were extremely steep regional potentiometric head gradients nearby. Due to this, the decision was made to develop Case C topography, where the river lies close to the edge of the floodplain.

4.2.8.2 MODFLOW versions and numerical stability

The Coonambidgal Formation has historically provided a challenge to groundwater models in the SA MDB. The low permeability of the unit, combined with unsaturated nature of the formation, has resulted in it being excluded from most regional models as it causes numerical instability. However, as the Goyder Floodplain Model focuses on floodplain processes and how to model them, it was necessary to include this unit as many of the key floodplain processes occur within it.

The Salinity Register Models were developed with MODFLOW2000 and was therefore selected as the code for the initial floodplain model. Initial simulations revealed that MODFLOW2000 was incapable of accurately simulating the wetting of the Coonambidgal Formation for losing stream conditions in steady-state. The re-wetting function within this code was employed, but to make the model converge, the re-wetting parameters used between base cases were inconsistent with each other. In addition, when this code was applied to the initial scenario runs, it was evident that the re-wetting function did not adequately represent changes in the Coonambidgal Formation under transient conditions (Figure 4-13).

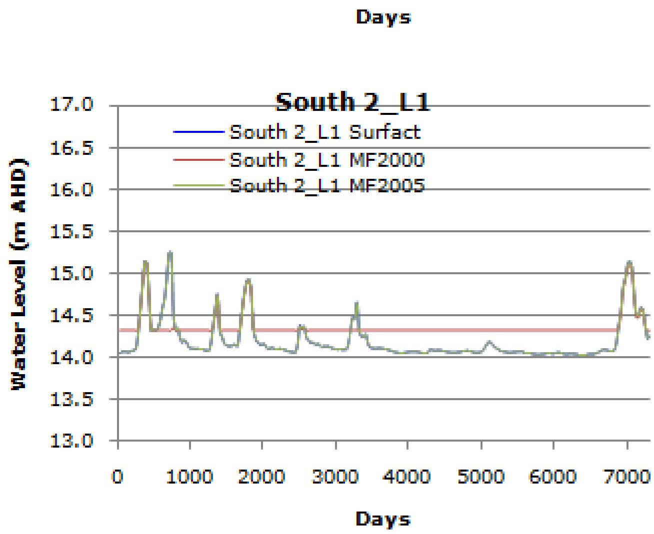


Figure 4-13 Comparison of MODFLOW variants when simulating wetting of the Coonambidgal.

Following that finding, both MODFLOW Surface (using the pseudo-soil option) and MODFLOW2005 with the NWT solver and UPW package were trialled, and a comparison of the different codes is shown in Figure 4-13. This figure shows a comparison of these codes for a simulation with a transient river level. Both MODFLOW Surfact and MODFLOW2005 with NWT are capable of representing fine temporal scale changes in the watertable. Both codes employ a 'residual saturation' method in addressing the issue of dry cells, where any cell that is calculated as being dry retains a minimal saturated thickness. This allows for the cell to re-wet without the convergence issues encountered in MODFLOW2000.

One other consideration that was made when selecting the code was the ability of the code to also have a compatible code to undertake solute transport modelling. Both MODFLOW Surfact and MODFLOW2005 NWT had this capacity, however MODFLOW Surfact used propriety code that was not available, whereas MODFLOW2005 used MT3DMS, a commonly used and available USGS code for solute transport. Therefore the code that was used to run the Goyder Floodplain Model was MODFLOW2005 (including the NWT solver and UPW package) for the flow modelling and MT3DMS for solute transport modelling.

4.2.8.3 Riverbed conductance and numerical instability

Riverbed conductance was tested in initial sensitivity runs and was found to have some impact on river fluxes, though not as much as the extinction depth of evapotranspiration. While the model was still being run in MODFLOW2000, higher riverbed conductances were found to cause a degree of model instability which required manipulating the re-wetting parameters to address. Hence lower riverbed conductances were selected.

4.3 Scenarios

The scenarios investigate the simulation of changes in river level, evapotranspiration, inundation and solute transport. The following outlines the construction of each of these scenarios, with the results discussed in the following section. Many additional scenarios were identified during the course of this project and these have been outlined in the recommendations for future work in Chapter 6.

4.3.1 River Scenarios

In the base case models, the river is represented by a boundary condition that is constant over time. This is similar to the method used in the Salinity Register models. The river scenarios assess the impact of time-varying river levels on the model. The specific time discretisations investigated were:

- Yearly average river levels
- Monthly average river levels
- Sub-monthly average river levels (adaptive time steps)

The river level for the model was based on the daily measured river level upstream and downstream of Lock 5 on the River Murray. River level elevation data for down and upstream lock 5 were obtained from WaterConnect (https://www.waterconnect.sa.gov.au), comprising daily recorded stage elevations for a 20 year period from 1/01/1992 to 31/12/2011 (Figure 4-14). The data sets for both up and downstream lock 5 included numerous gaps, some of which were greater than 7 consecutive days. A Python script was developed (Woods, 2015b; Appendix A) to read in each data set and fill the gaps in the datasets through linear interpolation to nearest known neighbour. The script is generic and may be used to fill gaps in any daily dataset; however, care should be taken if interpolating across a peak or trough in the dataset. The graphs produced by the script highlight where the filled data has been inserted (Figure 4-14).

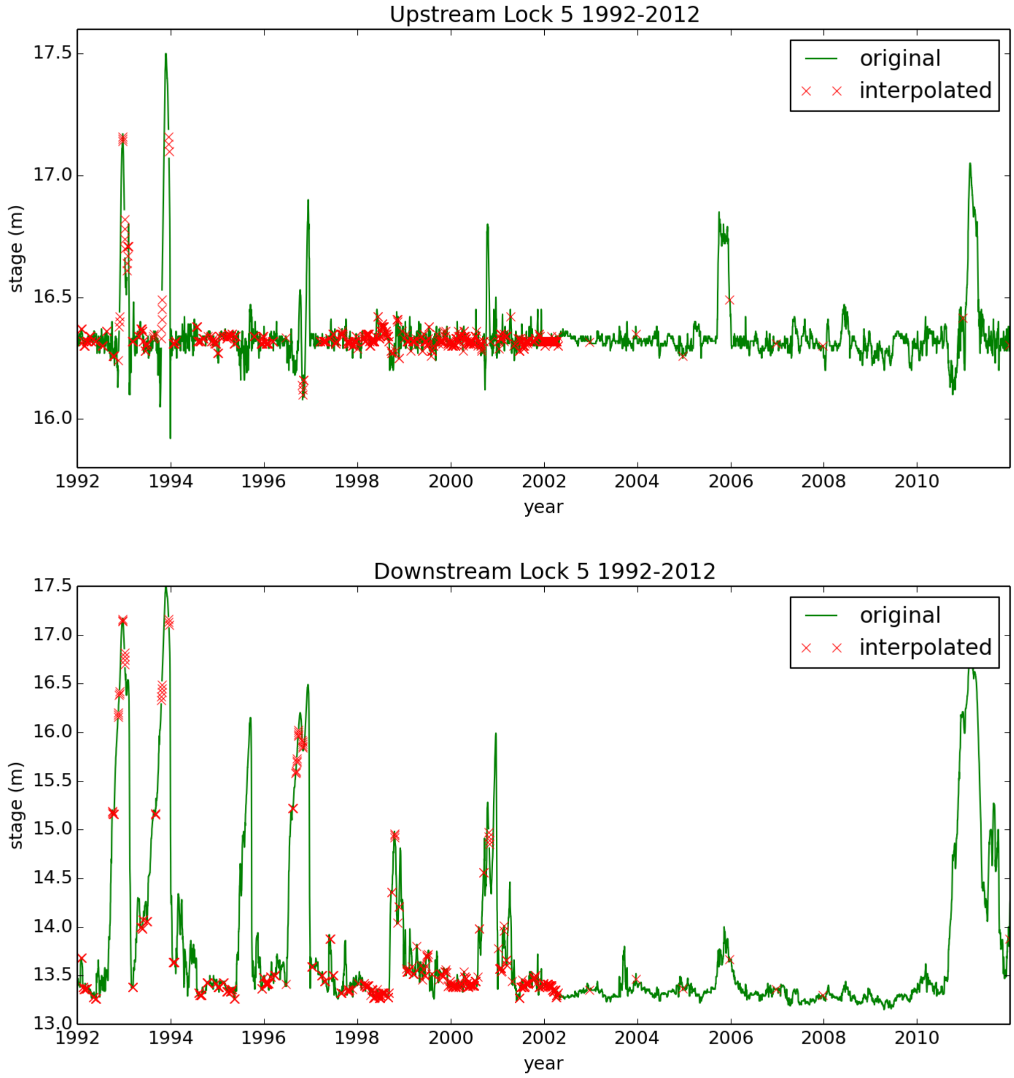


Figure 4-14 Measured River Level Data Upstream and Downstream of Lock 5

For the models with locks, this data was used directly, e.g. the average yearly, monthly or sub monthly level was calculated based on the measured data then directly input into the model. For the models without a lock, the river level was essentially synthetic so a different methodology was used.

The river level in the models without a lock was set as 2 m below the surface of the model. Given the topography of the model, this meant that the river level has a gradient consistent with the east-west gradient in the model. To capture the variation in river levels represented in the measured data and represent it in the model, the following methodology was followed:

- Calculate the difference between the daily measured river level and the pool level
- Calculate the average difference over the appropriate time period (yearly, monthly, sub monthly)
- Add that difference to the water levels in each river cell for each time step

This created a time-varying river level that maintained the river level gradient in the base models. The data that was used for this calculation was the measured river level downstream of Lock 5 as this contained the greatest variation in river level over time as compared to the water level upstream of Lock 5.

The aforementioned simulations do not include the effects from floods with respect to overbank land inundation. All model parameters for these simulations are constant with time apart from the river stage. The aim of these numerical experiments is to determine the effects of temporal discretisation on the floodplain hydrodynamics. To facilitate direct comparison, it is necessary to ensure that only within bank effects are modelled despite stage elevations exceeding adjacent land surface.

The scenarios for evaluating the impact of varying river levels are referred to as Scenario 1A and are outlined in Table 4-8.

Table 4-8 River Scenarios

Model Name	Stress Period Set up	With/without Lock	Case Based on
S1A_A_L_yr	yearly	With	A
S1A_A_L_mth	monthly	With	A
S1A_A_L_adp1	adaptive - coarse	With	A
S1A_A_L_adp2	adaptive - fine	With	A
S1A_A_NL_yr	yearly	Without	A
S1A_A_NL_mth	monthly	Without	A
S1A_A_NL_adp1	adaptive - coarse	Without	A
S1A_A_NL_adp2	adaptive - fine	Without	A
S1A_B_L_yr	yearly	With	B
S1A_B_L_mth	monthly	With	B
S1A_B_L_adp1	adaptive - coarse	With	B
S1A_B_L_adp2	adaptive - fine	With	B
S1A_B_NL_yr	yearly	Without	B
S1A_B_NL_mth	monthly	Without	B
S1A_B_NL_adp1	adaptive - coarse	Without	B
S1A_B_NL_adp2	adaptive - fine	Without	B
S1A_C_L_yr	yearly	With	C
S1A_C_L_mth	monthly	With	C
S1A_C_L_adp1	adaptive - coarse	With	C
S1A_C_L_adp2	adaptive - fine	With	C
S1A_C_NL_yr	yearly	Without	C
S1A_C_NL_mth	monthly	Without	C
S1A_C_NL_adp1	adaptive - coarse	Without	C
S1A_C_NL_adp2	adaptive - fine	Without	C

4.3.1.1 Reservoir cells

An alternative to the river package in MODFLOW is the reservoir package. It also functions as an internal head-dependent flux boundary with a flux limit and requires almost identical information to that of the river package. Differences between the two packages relate to their behaviour within a stress period. For a river cell, all parameters are set for a stress period and remain constant for every time step within. A reservoir cell has a starting stage elevation and a finishing stage elevation set for each stress period. The modelled stage elevation is then adjusted internally through linear interpolation for each time step within the stress period. River cells therefore follow a step function with respect to stress periods while the reservoir cells approach a piecewise linear function as numbers of time steps increase within a stress period. It should be noted that in a stress period with a single time step, there is no difference between the function of a river and reservoir cell if all input parameters are identical.

The River Murray in SA may be conceptualized as a sequence of reservoirs as the locks maintain a relatively flat pool level for each reach. The reservoir package is advantageous when editing temporal stage variation, as all cells assigned to a specific reservoir will vary as one. This approach does however, make it more difficult to use backwater curve data, which may have a much as a 1 m/km stage variation during periods of high flow rates. An approximation involves splitting a reach into 1 km segments, represented as separate reservoirs within the model. The suggested 1 km split is based on the approximate resolution of the backwater curve data.

Simulations implementing the reservoir cells as a replacement for the river do not simulate solute transport as MT3DMS does not officially support MODFLOW's reservoir package. It is nevertheless possible to manipulate the sink source mixing input file in MT3DMS so as to include the effects of the reservoir cells, however this was not attempted due to the time constraints placed on modelling.

The scenarios for evaluating the impact of using reservoir cells are referred to as Scenario 1B and are outlined in Table 4-9.

Table 4-9 Reservoir Scenarios

Model Names	Stress Period Set up	With/without Lock	Case Based on
S1B_A_L_yr	yearly	With	A
S1B_A_L_mth	monthly	With	A
S1B_A_L_adp	adaptive	With	A
S1B_B_L_yr	yearly	With	B
S1B_B_L_mth	monthly	With	B
S1B_B_L_adp	adaptive	With	B
S1B_C_L_yr	yearly	With	C
S1B_C_L_mth	monthly	With	C
S1B_C_L_adp	adaptive	With	C

4.3.1.2 Adaptive time step development for the river and reservoir boundaries

The river represents the most variable boundary condition of a floodplain model. Most groundwater models of the study area were developed to estimate fluxes to river from regional sources over yearly to decadal time spans, so river levels were held constant as "average" conditions. The current need for increased resolution of floodplain groundwater behaviour precludes static stage elevations and raises a question as to the range of spatial and temporal detail necessary for accurate simulation of floodplain hydrodynamics. Annual and monthly averages can be readily calculated for the river levels, but an adaptive method for discretising river levels in sub-monthly steps was also desired.

The intention of reducing the stress periods in the model to a sub-monthly level was to increase the temporal resolution when the river level changed dramatically. However, a consequence of increasing the stress periods is that model run time increases significantly. Therefore part of this investigation was to attempt to understand at what level of discretisation the point of diminishing returns was reached.

Creating an appropriate data set to investigate this within a groundwater model is a time intensive process and therefore a Python script (Woods, 2015b, Appendix A) was developed to streamline this process. The advantages of using a script included establishing a clear and repeatable methodology to base the discretisation on. In addition, multiple temporal discretisations could be assessed for the number of time steps they produced prior to being implemented in the model.

The script developed produces a yearly, monthly and sub-monthly adaptive step function. The adaptive step function is based on user-defined monthly splits for a specific range of differences between maximum and minimum river level elevation. For example, the user defines a series of rules in the following format, "If difference between maximum level and minimum level within a month is *x*, then split the month into *y* stress periods," where greater *x* leads to greater *y*. This effectively makes the river package mimic the behaviour of the reservoir package at the expense of increasing numbers of stress periods.

The generic rules used for the two adaptive river level scenarios were:

- Monthly maximum time step
- Two day minimum time step
- Time steps must be whole days

The specific rules, based on the difference between the maximum and minimum river levels within a month, for the adaptive river level scenarios are given in Table 4-10. The values were chosen such that the adaptive 1 scheme increases the discretisation for larger stage fluctuations should therefore provide greater detail for the largest flood peaks. The adaptive 2 scheme is set such that greater detail is captured in the smaller stage fluctuations.

Table 4-10 Temporal discretisation rules for river boundary scenarios

Within month river level difference (m)	Level of temporal discretisation (Month divided by)	
	Scenario - Adaptive 1	Scenario - Adaptive 2
0.1	-	2
0.3	-	3
0.5	2	5
0.6	3	-
0.7	5	14 (~2 day stress periods)
0.8	14 (~2 day stress periods)	-

An additional Python script was developed to obtain the piecewise linear function necessary for the reservoir package. The initial approach was to tag peaks and troughs in the data set and use these for the start and end stage in each stress period. Numerous small stage fluctuations in the data set, lead to increasingly large numbers of one or two day stress periods. To minimize the number of stress periods required, the data was processed through convolution with a Gaussian kernel. Smoothing of the data set with a large window or high pass kernel (10 days) resulted in unacceptable reduction of the flood peak elevations. Consequently, the data set was split via a threshold elevation. Days in the dataset where the stage elevation exceeded the threshold were replaced with the threshold stage elevation value. A high pass convolution was then performed on the modified dataset. This resulted in significant smoothing of the smaller scale fluctuations in river stage and is tantamount to the adaptive 1 scheme for the river cells, which aims to provide greater detail for large stage fluctuations. Days where stage height exceeded threshold were then set back to their original values thus maintaining unmodified flood peaks within the working data set. The peaks, troughs and inflection points of the resulting function were then tagged to provide the start and end stage for each reservoir stress period. Inflection points were determined via local second derivatives using a centred finite difference approach (Figure 4-15).

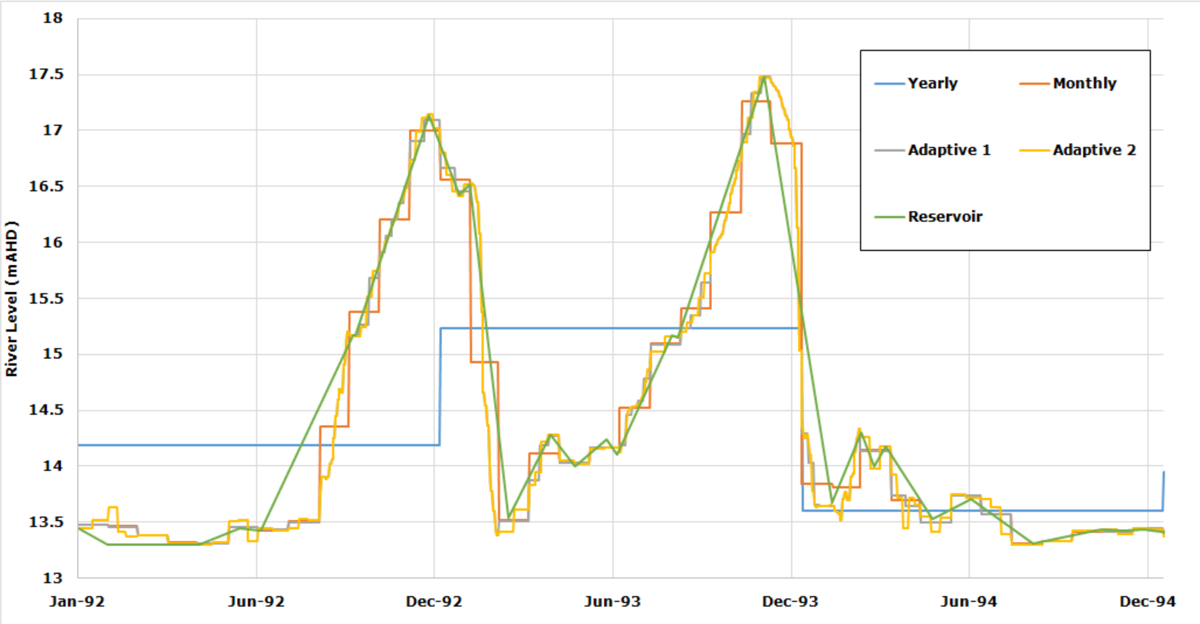


Figure 4-15 River levels at different timesteps

4.3.1.3 Changes in cell size

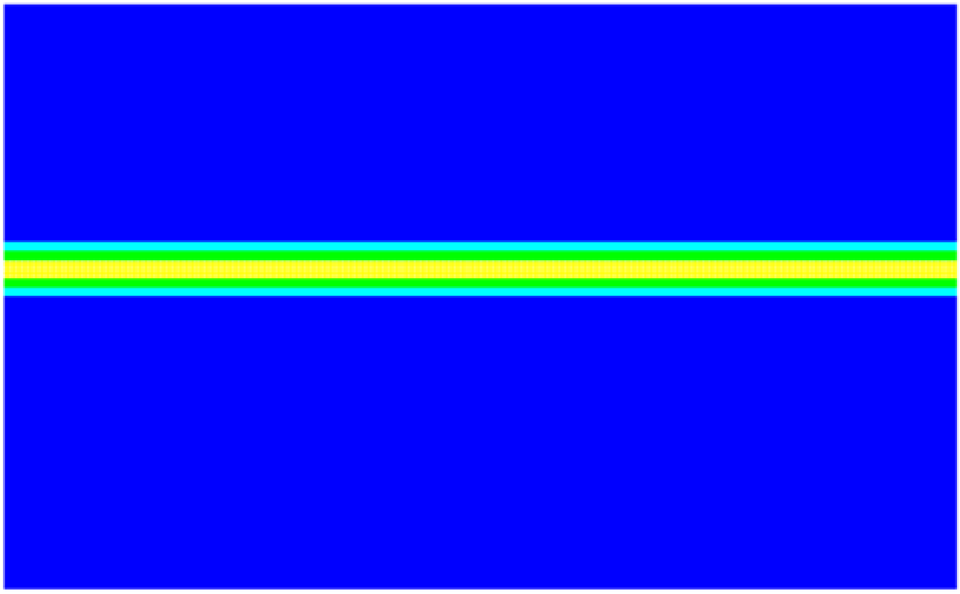
MODFLOW-USG was used to investigate the impact of changes in cell size in the vicinity of the river. MODFLOW-USG is a version of MODFLOW which allows the use of an unstructured grid, allowing grid refinement to occur in specific areas of interest, without impacting on the model geometry outside of these areas. Scenario 1A, both with a lock or without a lock, was used as a basis for this scenario. Model cells were refined using quadtree refinement on and near the river. Quadtree refinement progressively subdivides a model cell into 2^(quadtree_level), for some user-specified quadtree level between 2 and 8. For example, a quadtree level of 2 means a cell will be subdivided into 4 equal cells, while a quadtree level of 4 will result in a cell being subdivided into 16 equal cells.

For the river scenario, a maximum quadtree level of 4 was applied to the cells hosting the river (yellow zone in the figure below) resulting in 15.625m x 15.625m cells. Cells either side of the river were refined lower quadtree levels resulting in a smoothly refined grid. Figure 4-16 shows the quadtree levels used in the model.

The river cell hydraulic conductivity parameter was modified to result in a river cell conductance that was equivalent to the MODFLOW2005 version of the model. The models were run using the UPW package to maintain consistency with the MODFLOW2005 models run previously and to avoid potential convergence and stability problems.

Results from these scenarios show that the modelled water levels are consistent with the equivalent MODFLOW2005 model. Mass balance calculations indicate that the modelled flux to the river is higher than that calculated in the MODFLOW2005 model.

Due to time constraints, these results were not investigated further or for models constructed for other scenarios. However, further developing this work has been included as a recommendation in Chapter 6.



Blue = 125x125, Light blue = 62.5x62.5, Green = 31.25x31.25, Yellow = 15.625x15.625

Figure 4-16 Grid refinement around river

4.3.2 Evapotranspiration Scenarios

The evapotranspiration scenarios were designed to investigate how various components of evapotranspiration are represented within groundwater models and the impact on how the model represents floodplain systems. As such, the scenarios focus on what can be varied within a groundwater model and less on the science underpinning the values used. This is acknowledged as a limitation of this work and the recommendations in Chapter 6 attempt to highlight further work that may be undertaken to address this issue.

Evapotranspiration, as with river levels, is commonly modelled as a value that does not change with time. While there are some models that do vary evapotranspiration in space, it is common for groundwater models of the study area to have one or two evapotranspiration zones held constant over time. In addition, most MODFLOW models use the MODFLOW evapotranspiration package, which employs a linear decrease in evapotranspiration rate with depth to a nominated extinction depth. The scenarios developed were designed to investigate the impact of this methodology.

The scenarios are outlined in Table 4-11. The evapotranspiration rate used in the model was based on daily FAO56 evapotranspiration data from the Loxton Research Centre (Station No. 24024). This was averaged over the appropriate timescale (yearly or monthly) and used as an input to the model. Extinction depth was based on the depth used in the base model (2.5 m) and varied during scenarios C and D to investigate the sensitivity of the model output to this parameter.

As the model is synthetic, a synthetic spatial distribution was developed to represent how changes in soil or vegetation might impact on evapotranspiration. As discussed above, the science to support this distribution was not investigated in great detail, however an attempt was made to correlate the synthetic distribution with real world data. This distribution (Figure 4-18) was based on soil and vegetation maps from various floodplains in the border to Lock 3 area (AWE, 2011; AWE, 2012a; AWE, 2012b). The soil maps showed that clays and large trees were more prevalent near surface water features, which is a trend that was mimicked in the synthetic distribution developed. In addition the extinction depth was varied spatially as well, as specified in Figure 4-18. This was done as part of a sensitivity analysis and not based on any real world data.

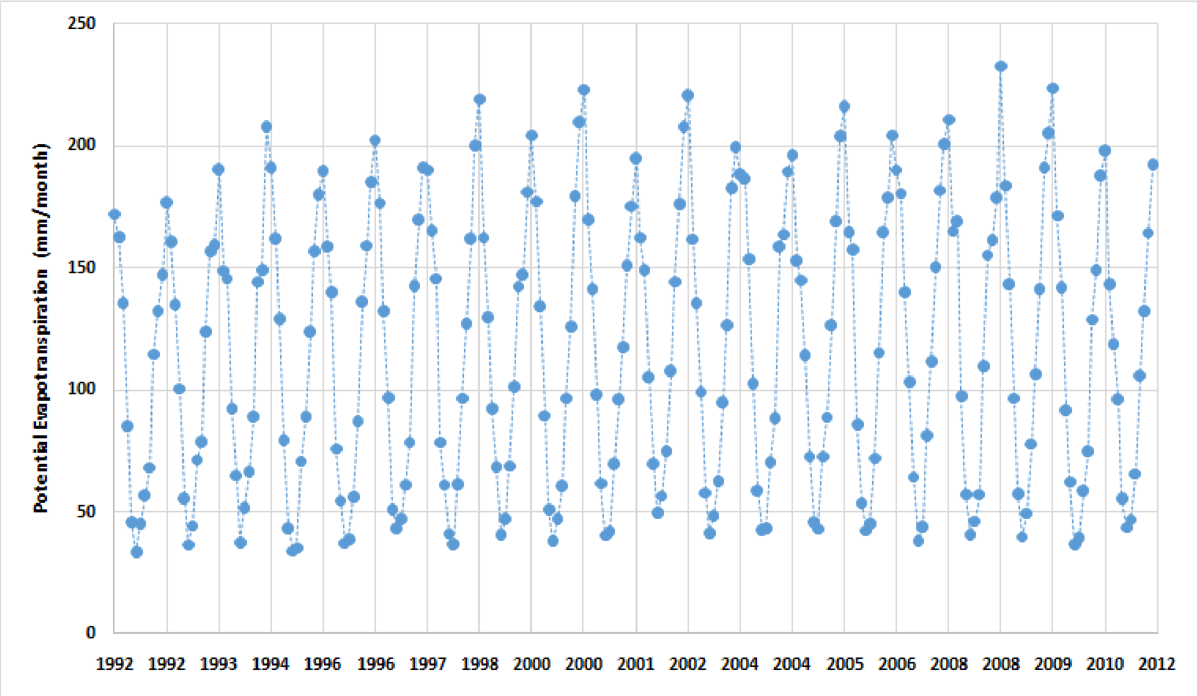


Figure 4-17 Potential evapotranspiration varies over time

Table 4-11 Evapotranspiration Scenario Outline

Scenario name	Time scale of variation	Ext Depth (m)	Spatial variation	ETS Package
Scenario 2A	Yearly	2.5	No	No
Scenario 2B	Monthly	2.5	No	No
Scenario 2C	Monthly	2	No	No
Scenario 2D	Monthly	3	No	No
Scenario 2E	Monthly	2.5	Yes - ET rate only	No
Scenario 2F	Monthly	1.5 - 3	Yes - ET rate and depth	No
Scenario 2G	Monthly	2.5	No	Yes

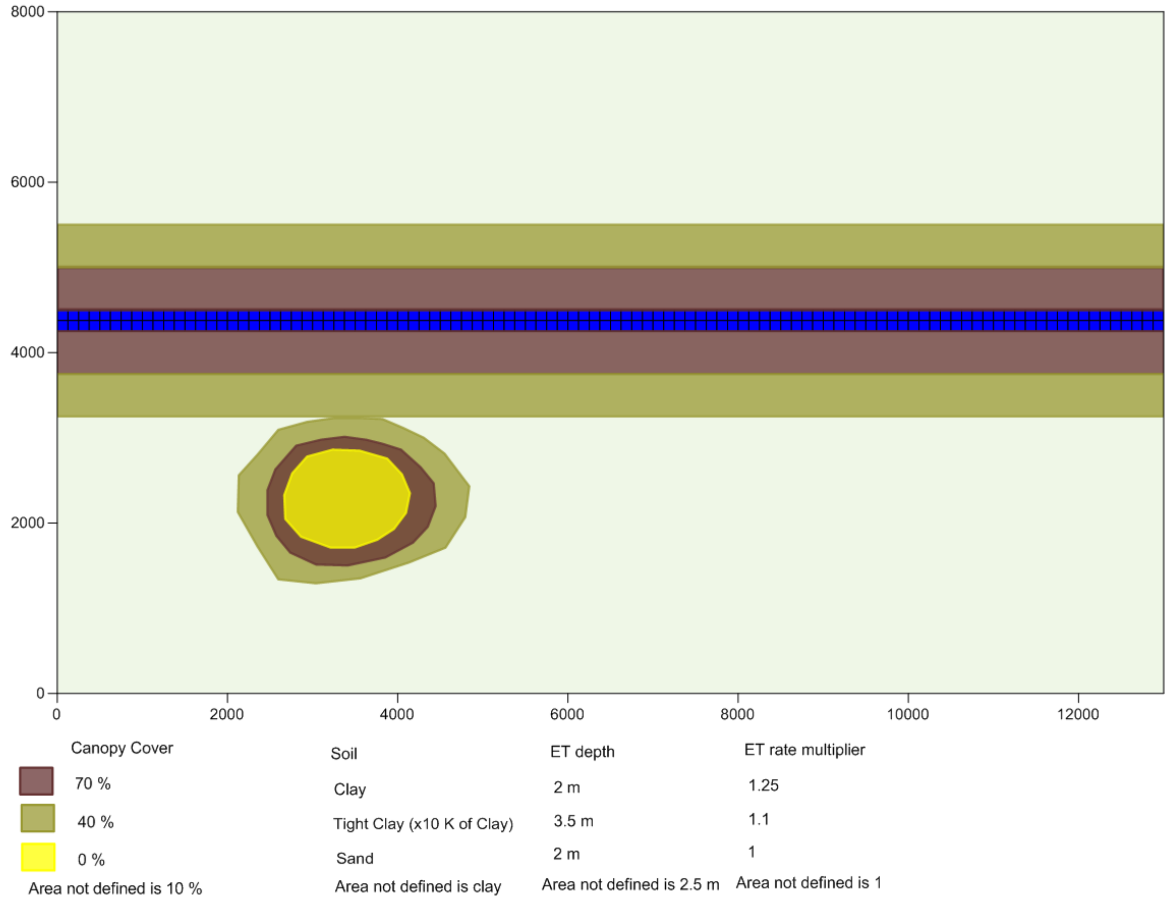


Figure 4-18 Spatial distribution of evapotranspiration

In addition to the changes made within the MODFLOW ET package (scenarios A to F), a scenario was undertaken to investigate the impact of using a non-linear relationship evapotranspiration and depth. This was done through the ETS package and based on a series of equations for this non-linear relationship given in Shah *et al.* (2007). This paper provides non-linear relationships between extinction rate and depth for a range of soil types and vegetation. The curve selected for use in Scenario G was based on the sandy clay with grass cover parameters outlined in this paper. The justification for this selection was that the soil type needed to be within the clay range and that the extinction depth (when the evapotranspiration is essentially zero) should be similar to the depth used in the base model. The second reason was so that results between the scenarios could be more directly compared without the complication of a different extinction depth. The curve used in the model is given in Figure 4-19.

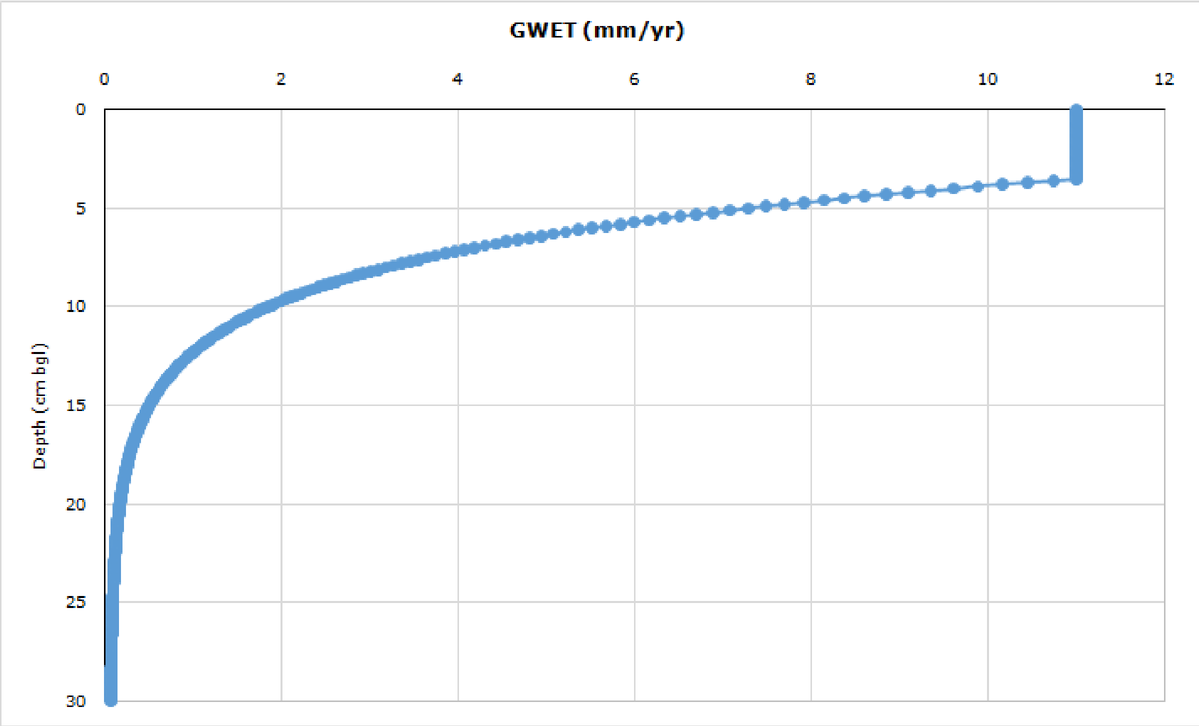


Figure 4-19 ETS function used in Scenario 2G

4.3.3 Inundation Scenarios

The initial conceptualisation of the inundation scenarios were to investigate not only the impact of inundation, but also variations in how the recharge due to inundation is calculated as well as the use of reservoir cells. Due to time constraints, this was not achievable, however some inundation scenarios were able to be undertaken to investigate the impact of including inundation.

Inundation was implemented in the model as recharge to groundwater. This is consistent with the methodology used in the Chowilla Groundwater Model (RPS Aquaterra, 2012). The rates used were also consistent with the Chowilla Groundwater model. An outline of the scenarios undertaken is given in Table 4-12.

Table 4-12 Inundation Scenarios

Scenario name	Stress Periods	Inundation Recharge Rate (mm/day)	Area inundated	Spatial variation
Scenario 3A	Monthly	1	Floodplain + Wetland	No
Scenario 3B	Monthly	1	Wetland only	No
Scenario 3C	Monthly	0.5 - 2	Floodplain + Wetland	Yes

As the model was built with a wetland area, a separate surface water balance model was developed to determine the area inundated in the wetland, due to flooding. This was a non-iterative, daily spreadsheet model that accounted for the following inflows and outflows:

- Inflows
    - Flooding – based on river level
    - Rainfall – based on daily measured rainfall at the Loxton Research Centre
- Outflows
    - Evapotranspiration - based on daily FA056 evapotranspiration at the Loxton Research Centre

- Overflow – calculated as water in excess of the volume of the wetland
- Loss to groundwater – based on 1mm/day recharge when inundated

These inflows and outflows were calculated based on the area of the wetland inundated at any given point in time. Flow into the wetland was restricted based on the equation for flow over a broad crested weir. This was done to prevent the wetland from instantaneously filling in an unrealistic manner. In addition, for the scenarios with the lock, the assumption was made that any water above bank level upstream of the lock would not cause inundation, but be redirected to the wetland. This allowed for some difference between the floodplain and wetland inundation which may be representative of an environmental watering event.

Inundated areas outside of the wetland were based on topography. The average river level for a month was calculated and areas lower than that contour were inundated. An example of this is shown in Figure 4-20 and Figure 4-21 below.

A scenario was run where the wetland was the only component inundated. This scenario was designed to be consistent with the surface water model Source and was intended to be compared to a similar scenario run in that model.

As well as inundating with a flat rate of 1 mm/day, a third scenario was set up which was variably-inundated based on the distribution used in the evapotranspiration scenarios. The area inundated was overlaid on this distribution and the rates of the different areas applied accordingly (Figure 4-22).

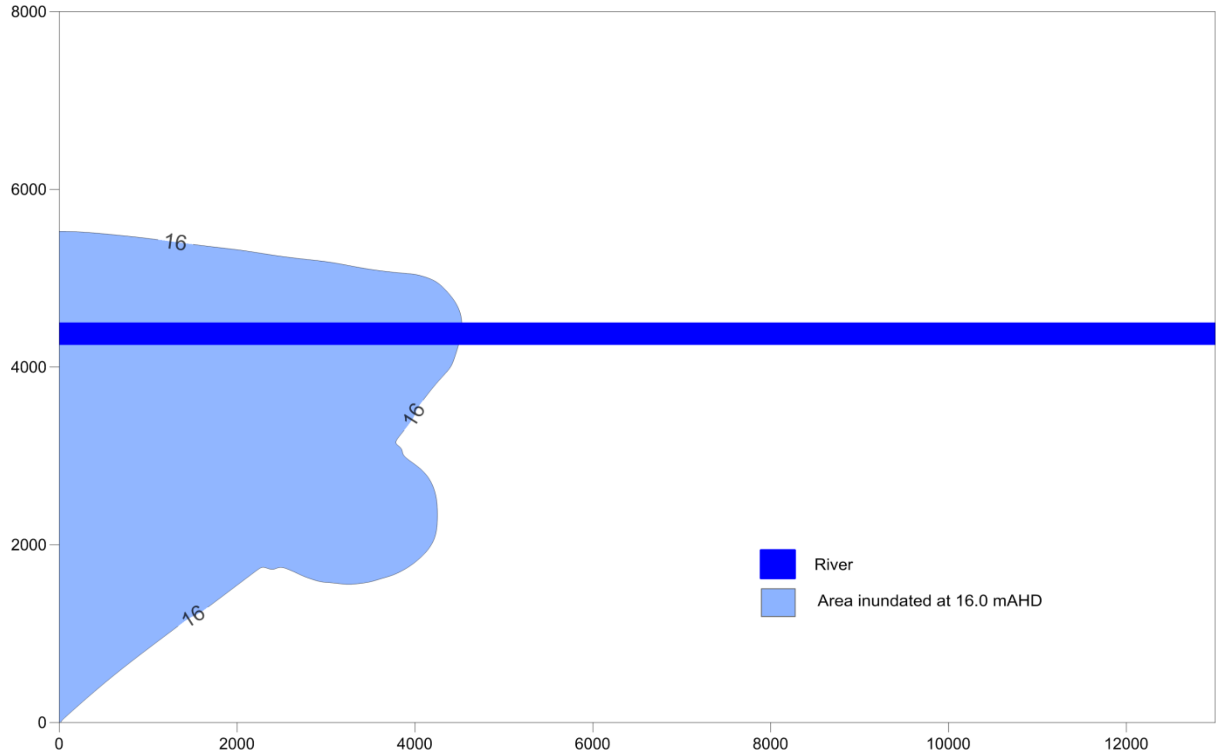


Figure 4-20 Inundation at 16.0 m AHD for Case A with lock

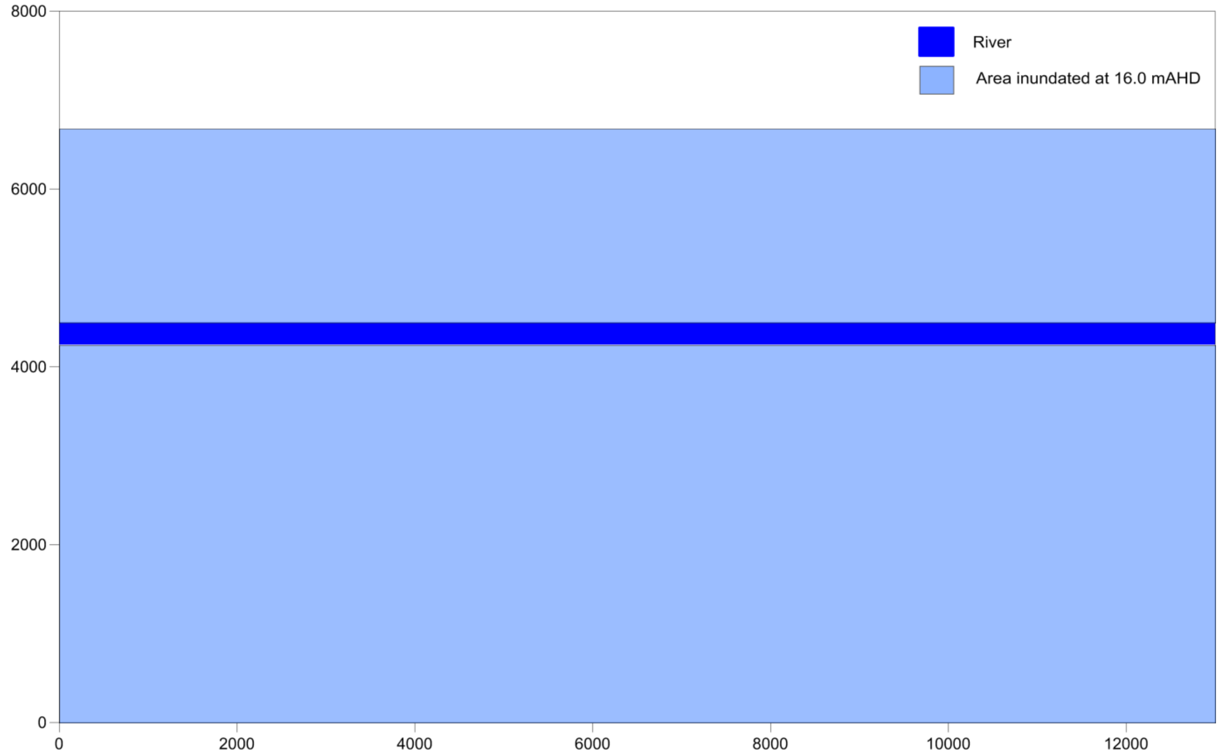


Figure 4-21 Inundation at 16.0 m AHD for Case A without lock

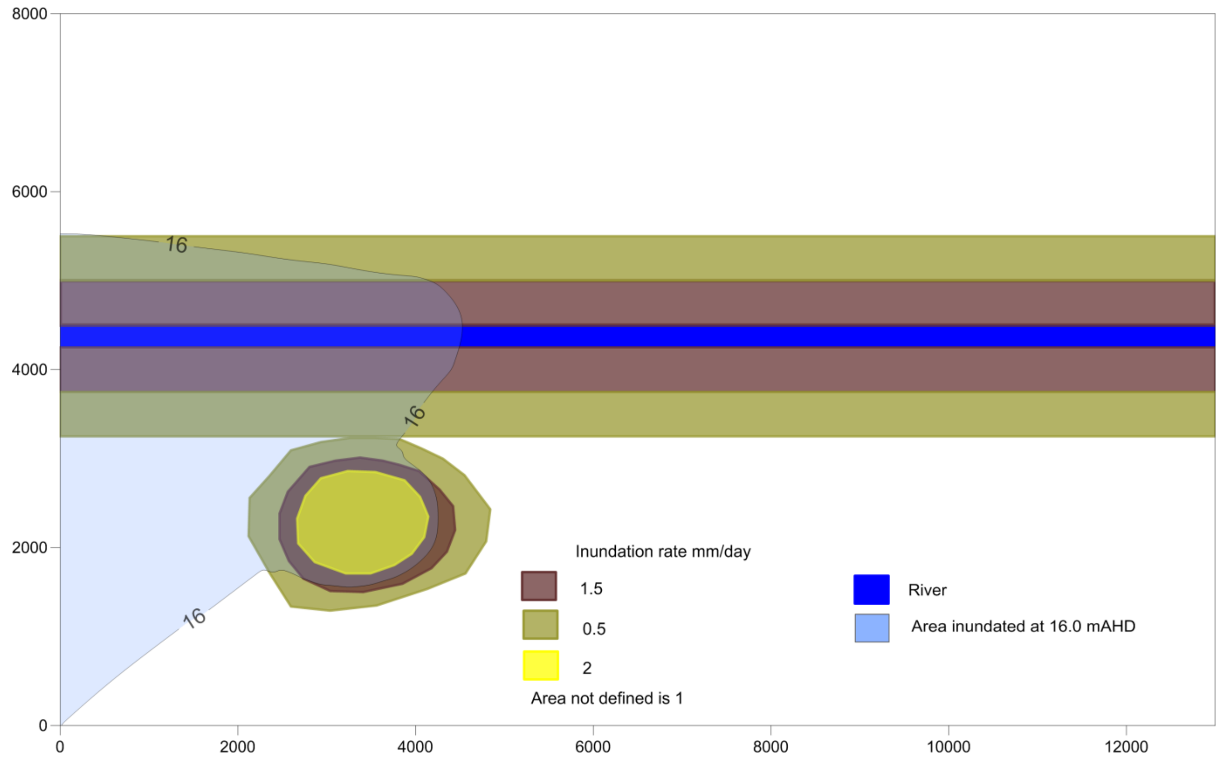


Figure 4-22 Inundation at 16.0 m AHD for Case A with lock overlain on spatial distribution of inundation rate

4.4 Scenario Results and Discussion

As outlined in the previous sections, six base models were constructed (Case A, B and C, with and without a lock). Each scenario was investigated for each of these base models (unless technically unviable, see Section 4.3.1.1). A suite of results have been produced for each of these runs, including:

- Hydrographs
- Water balance components over time, specifically:
  - Evapotranspiration
  - Flux into and out of the northern half of the river
  - Flux into and out of the southern half of the river
  - Horizontal flux to the cells hosting the river (north and south)
  - Vertical flux to the cells hosting the river (north and south)
  - Flux between the Coonambidgal Formation (layer 1) and the Monoman Formation (layer 2)
  - Flux between the regional aquitard (layer 3) and the Murray Group (layer 4)
  - Flux into the Monoman Formation through the general head boundaries
  - Flux into the Murray Group through the general head boundaries
  - Cumulative flux to river

These results have been compiled in Appendix C of Woods (2105b). The hydrographs are based on a series of 16 observation wells in layers 1 and 2, shown in Figure 4-23 (without lock setup) and Figure 4-24 (with lock setup).

For consistency and brevity, the following discussion of results will focus on the following model outputs:

- Cumulative salt load to river
  - For flow-only models salt load to river will be calculated as flux to river multiplied by 25,000 mg/l (assumed salinity of the Monoman Formation)
  - For flow and solute transport models, this is calculated as flux to river multiplied by groundwater salinity at the river
- Losing or gaining river changes over time
  - This is calculated as the difference between river inflows/outflows for the entire length of the river and is not representative of smaller scale spatial changes
- Water balance and salt balance at December 1996
  - For yearly models this will be 1996
  - For sub monthly models this will be the end of December 1996
- Potentiometric surfaces and salinity contours at December 1996
  - For yearly models this will be 1996
  - For sub monthly models this will be the end of December 1996
- Spatial evapotranspiration and river boundary contours at December 1996
  - For yearly models this will be 1996
  - For sub monthly models this will be the end of December 1996
- Water balance and salt balance at December 2006
  - For yearly models this will be 2006
  - For sub monthly models this will be the end of December 2006
- Potentiometric surfaces and salinity contours at December 2006
  - For yearly models this will be 2006
  - For sub monthly models this will be the end of December 2006
- Spatial evapotranspiration and river boundary contours at December 2006
  - For yearly models this will be 2006
  - For sub monthly models this will be the end of December 2006

These times were chosen as 1996 was a high flow year, preceded by numerous high flow years, while 2006 was a dry year, preceded by numerous dry years. While some of these figures will be included in the discussion below, a full set of these figures in included in Appendix C of Woods (2015b).

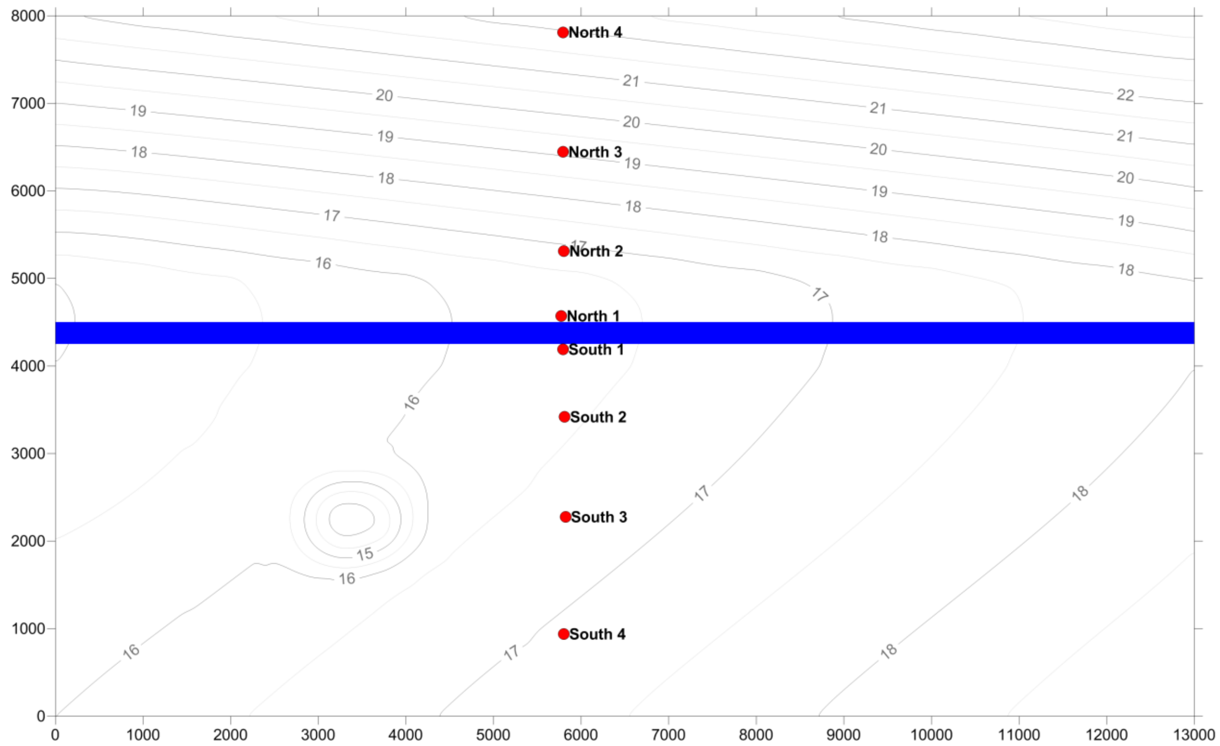


Figure 4-23 Observation bores in models without a lock

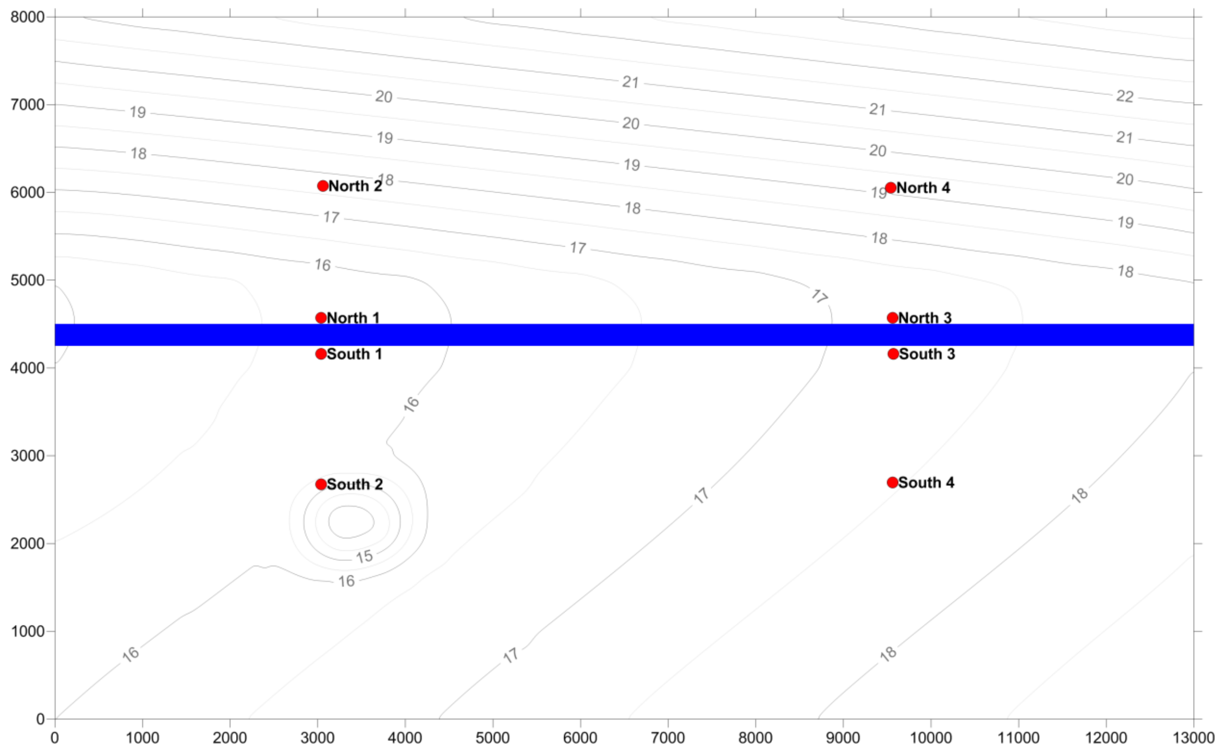


Figure 4-24 Observation bores in models with a lock

4.4.1 Scenario 1 - River Scenario

4.4.1.1 River Scenarios without Locks

The rivers in scenarios where a lock is absent have a sloping stage elevation that mimics the slope of the topography. The downstream lock 5 stage elevation dataset is used for the entire reach, that is, the entire reach in the model fluctuates as one. Increased losses are expected from the river to the aquifer during floods as a consequence of stage increase. A greater refinement of the stress period will therefore increase the magnitude of the loss and also have a greater impact on the adjacent floodplain hydrodynamics.

Potentiometric head in the floodplain aquifer is not strongly altered by the choice of temporal discretisation of river levels, provided monthly or adaptive stress periods are used. Figure 4-25 provides the results for Case A but the other cases have similar results.

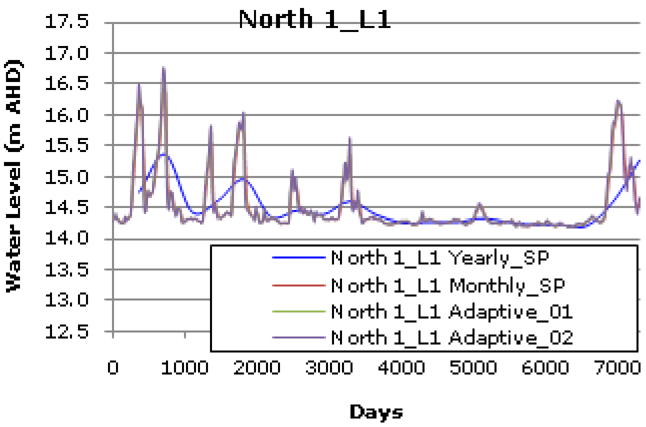


Figure 4-25 Hydrograph for Case A without lock

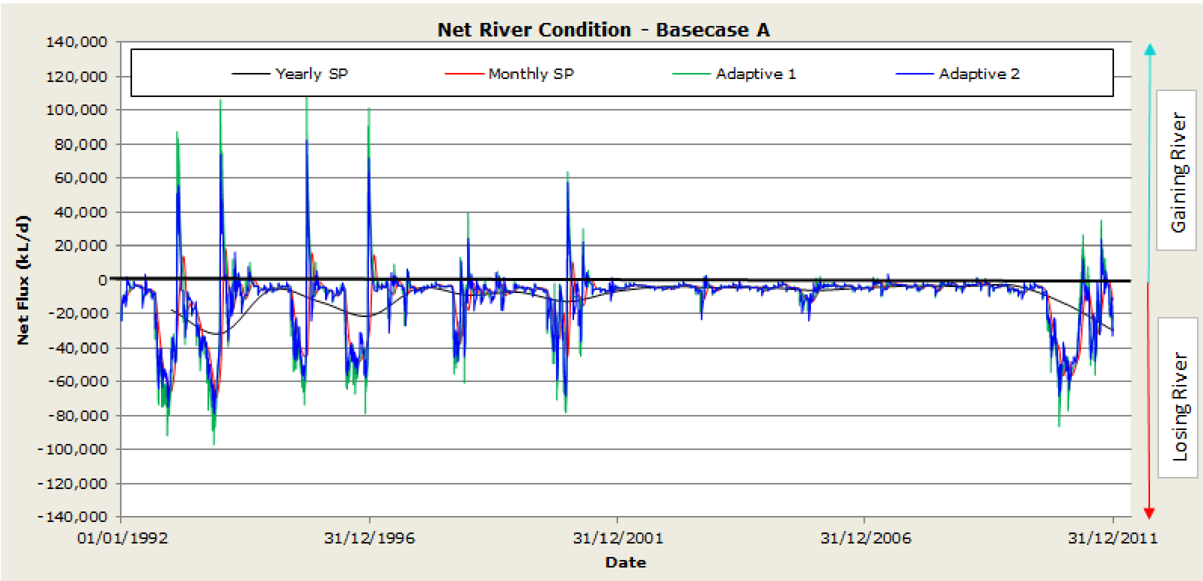


Figure 4-26 Net river condition for model A without a lock.

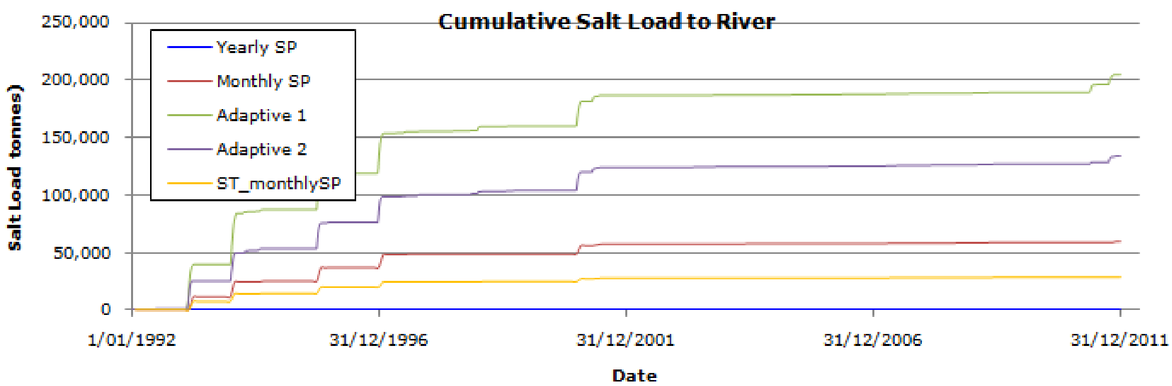
However, the time discretisation affects flux between the river and groundwater. Figure 4-26 shows the net river condition for Case A, with other cases showing similar results. Periods of increased loss from rivers coincide with high river levels.

A yearly discretisation can result in a river that always has net losing stream conditions, even during flood recession periods. The temporal signal of river level change is, however, captured well by monthly and adaptive stress period schemes, but the magnitude of the flux variation is better captured with an adaptive approach. This implies that a monthly discretisation scheme may not be adequate for capturing fluxes attributed to short-term events. In this instance, an adaptive scheme that increasingly discretises large changes in river stage (adaptive 1 scenario) results in significantly greater fluxes.

The effect of refining temporal discretization is shown most clearly in a comparison of the cumulative salt loads to river. Figure 4-27 shows the results Scenario A without a lock. Yearly discretisation yields a cumulative salt load that has a near-constant trend that is only partially modified by changing river levels. The other scenarios show rapid salt accessions in brief periods after a flood, with slower salt accessions in inter-flood and flood periods. While the trends are similar for the monthly and adaptive discretisations, the magnitude differs, with the adaptive methods providing the largest flux.

Figure 4-27 also shows how cumulative salt load depends on whether a constant-concentration approximation is used or a solute transport calculation. The cumulative salt load for this simulation as calculated using a solute transport simulation is approximately half of what is estimated when groundwater flux to the river is multiplied by a constant concentration. In this instance solute transport simulations appear to consistently provide reduced accessions unless the yearly mean discretisation scheme eliminates all gaining episodes for the entire 20 year period.

It should be noted that the magnitude of the fluxes are likely underestimated for these simulations due to the absence of overbank flood inundation during periods of elevated river stage. All the simulations in this scenario assume within bank stage fluctuation irrespective of adjacent land surface elevation. It is also worth noting that despite the steady state gaining conditions observed in base model C, it is predominantly a losing river reach with



large gain fluxes following floods.

Figure 4-27 Cumulative salt load for model A without lock.

December 1996 is considered a wet period with downstream lock 5 river stage receding from a flood (

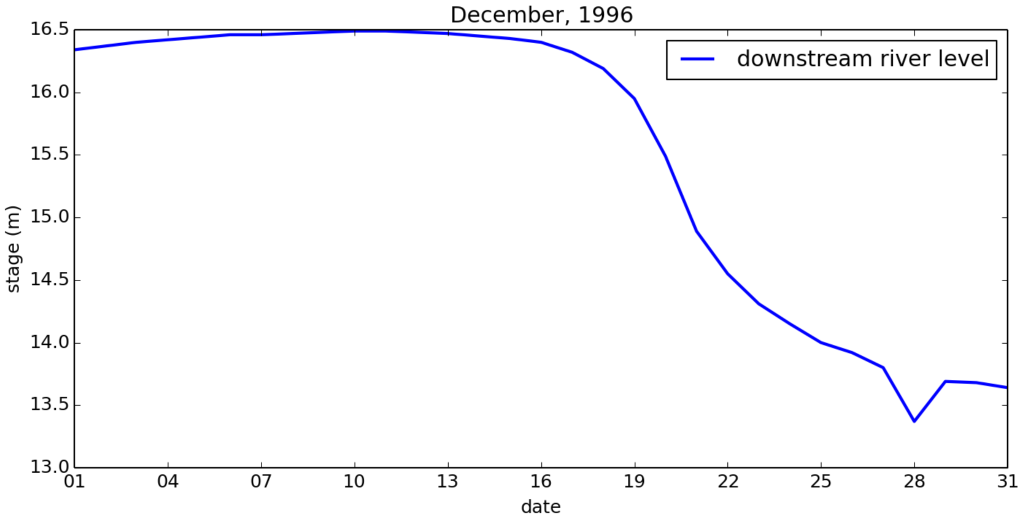


Figure 4-28). For this instance, the adaptive schemes capture the significant change to river stage while yearly mean and monthly do not. An examination of the water balance at this point in the simulation (Figure 4-29) further reinforces the effects of increased temporal discretisation with significant differences observed in fluxes for adaptive stress period schemes. Adaptive methods capture the gaining conditions associated with a flood recession in the later part of the month while yearly and monthly schemes display net losing river conditions.

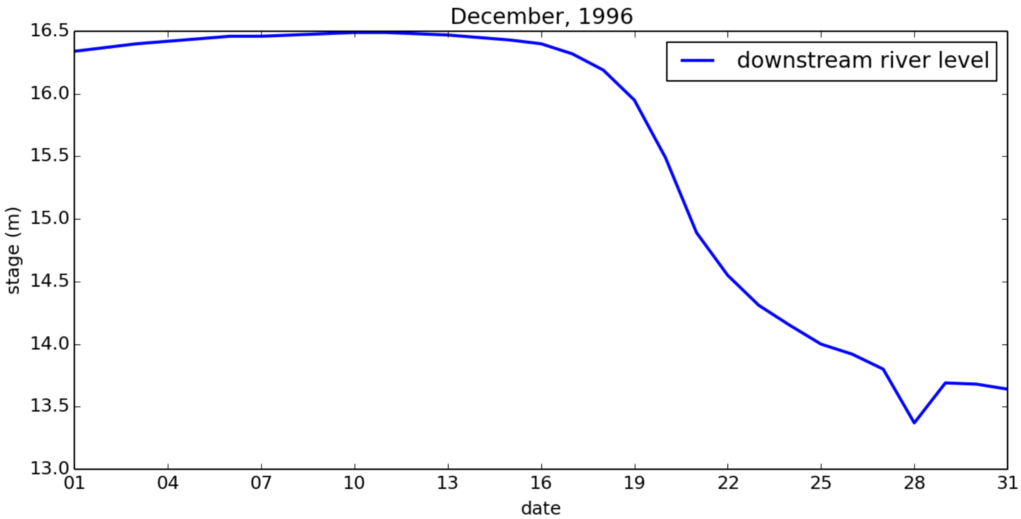


Figure 4-28 Downstream Lock 5 stage elevation for December 1996.

The mass balance figure (Figure 4-29) and others that appear throughout this chapter serve to identify the sources and sinks of water in the model. A flux labelled as in is effectively a source for the groundwater while those labelled out are sinks. For example, river out identifies water leaving the groundwater and entering the river i.e., a gained flux to the river or alternatively, a loss from the aquifer. By combining river in and out we get the net condition for the river in the model. **storage** is the change in water stored in the aquifer, **ch** refers to flows in and out of constant head boundaries (not used in the model), **riv** refers to flows in an out of the river, **et** is evapotranspiration, and **ghb** are flows in and out of general head boundaries.

Here, adaptive methods are indicating a net gaining river condition while the yearly and monthly time step schemes for the same time period indicate losing conditions.

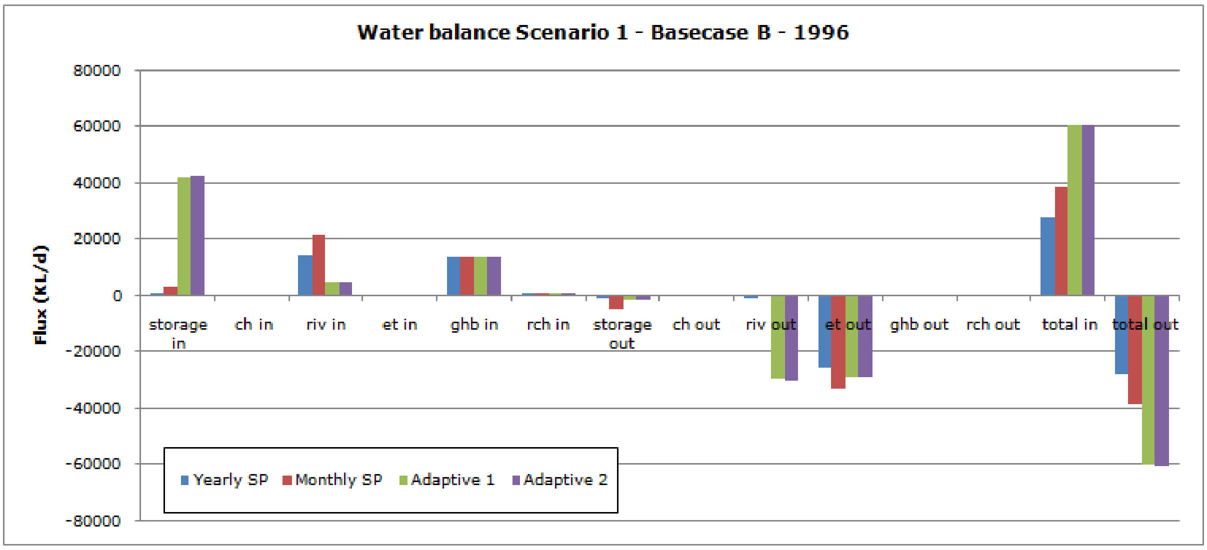


Figure 4-29 Water balance for Scenario 1 model B without lock December 1996.

4.4.1.2 River Scenarios with Locks

The hydraulic impact of a lock (i.e. a change in weir pool level) on the floodplain results in a freshwater flush zone where the river loses water into the aquifer along the upper reach and gains groundwater along the lower reach near the lock. Consequently, a significant groundwater flux volume is gained on the lower lock reach that is a closer match to river salinity than the surrounding aquifer. Adopting the approach where gained volumes are assigned a set concentration over-estimates salt accession to the river. This is evidenced in all 3 models with locks when comparing results between simulations with and without solute transport.

The impact on the floodplain flow regime attributed to a lock is best demonstrated by comparison with an identical model that excludes it. The cumulative solute flux for a scenario with a lock is significantly greater than that calculated for an equivalent scenario without a lock (compare Figure 4-27 with Figure 4-30). The impact of changing river levels is proportionally weaker, so the cumulative flux increases generally linearly with time, rather than in the stepped fashion seen for the scenarios without a lock.

The temporal discretisation of river levels affects scenarios with locks similarly to those without. Monthly and adaptive stress periods are sufficient to simulate changes in watertable (see hydrographs in Appendix C). However, net river condition and cumulative salt load is sensitive to the discretisation.

In models with locks and the adaptive 1 method, which is more suited to large stage variation, displays a reduced magnitude of net fluxes in comparison to scenarios without a lock. This distinction is most likely due to the damped upstream lock 5 stage variation over half the river implemented in models with a lock. The rivers in the models without locks are all simulated entirely using downstream lock 5 dataset, which has significantly larger stage variation and therefore more suited to the adaptive 1 scheme.

Figure 4-30 shows the cumulative salt load for Scenario A with Lock. The relatively constant rate of salt accession also demonstrates that the prevailing flow regime dominates and floods appear to have minimal impact on the total salt gained by the river in the vicinity of a lock.

Model water balances for December 1996 in all scenario 1 simulations show very similar results with the models with a lock displaying slightly lower fluxes than the identical model without a lock. For the sake of brevity these will not be displayed here (see Appendix C in Woods (2015b) for comparisons).

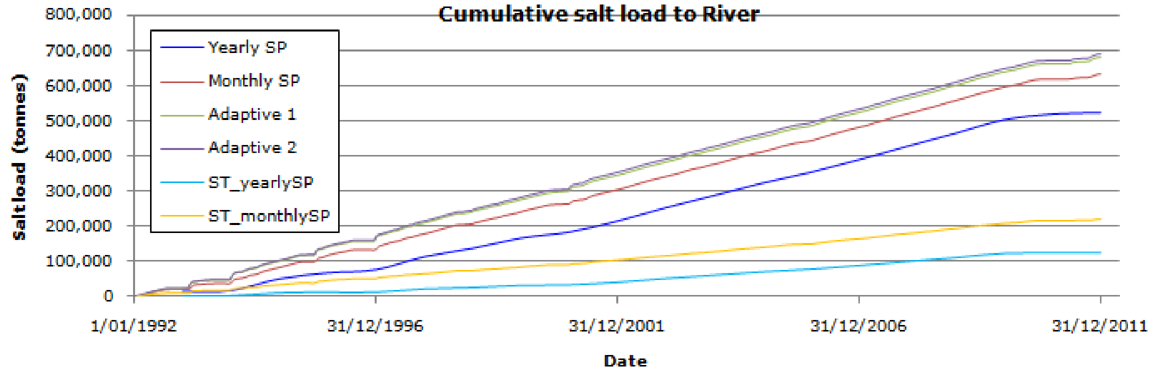


Figure 4-30 Cumulative salt load to river for model A with lock. Solute transport results are labelled with ST prefix. See appendix for other modelled results.

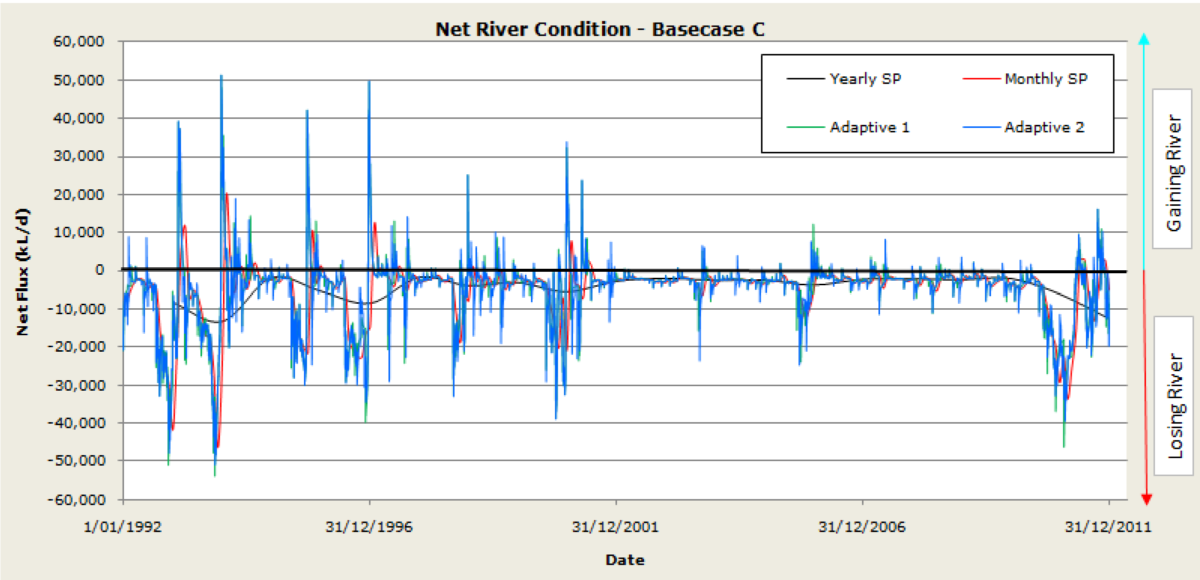
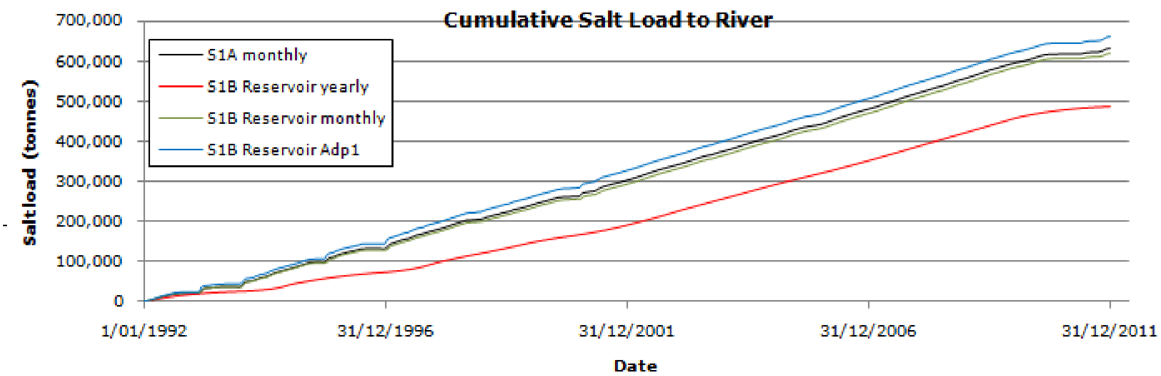


Figure 4-31 Net river condition for model C with lock displaying large rapid gaining fluxes.

4.4.1.3 Reservoir Scenarios

Reservoir cells were used to replace river cells in the models with locks. Reservoir cells are more difficult to implement where sloping stages of backwater curve are needed to accurately simulate the river. Another limitation of reservoir cells is the absence of official support for the package in MT3DMS, precluding its use in the solute transport simulations.

There is minimal difference between the use of the river and reservoir cells (Figure 4-32). A comparison with the monthly mean river cell approach produces nearly identical cumulative salt load values. The adaptively-discretised reservoir method displays some increase to the total flux but the difference less than the adaptive scheme used



with the river cell models. The yearly stress period scheme also produces similar accessions to that of the river cell. The net river condition for model C with a lock and reservoir cells (Figure 4-33) shows decreased net flux variation in comparison to the same model with river cells.

Figure 4-32 Cumulative salt load to River with reservoir cell model

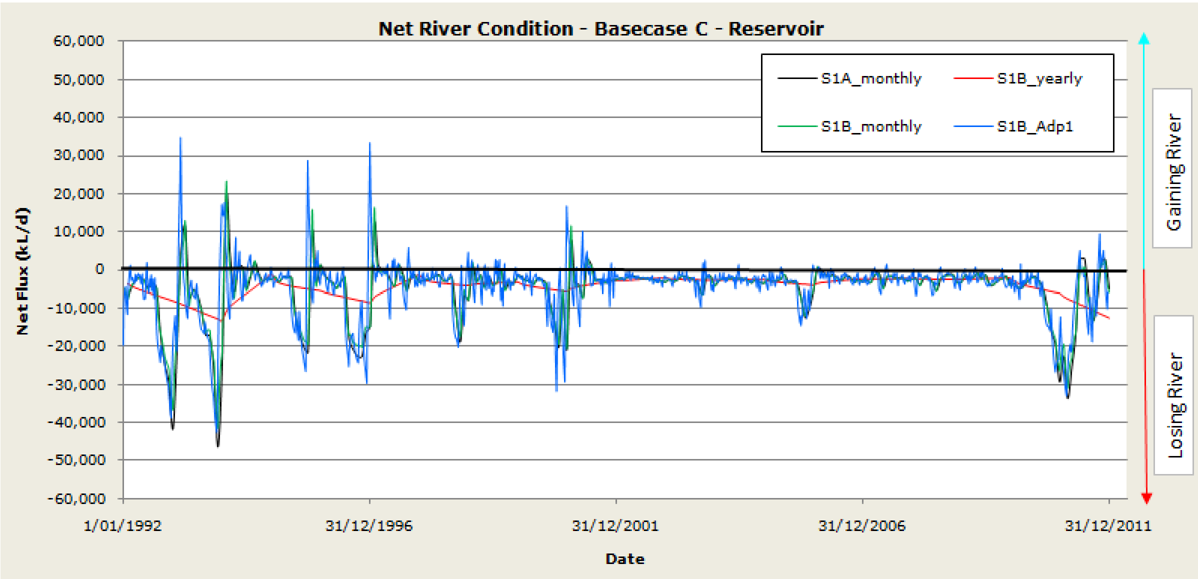


Figure 4-33 Net river condition for model C with a lock and reservoir cells

4.4.1.4 Discussion

It should be noted that magnitudes of stage fluctuation are fixed along each reach in all simulations. In reality river stage fluctuation is damped as we traverse further downstream. This is evident in the upstream lock 5 dataset (Figure 4-14), which has significantly less stage fluctuation for the same period of time. Backwater curve data for the various lock reaches of the Murray in SA give some indication as to the reduction in stage fluctuation along a reach. It is unclear how significant the effect of a river modelled with a backwater curve has on the floodplain.

The adaptive methods combined with the Python script provide a relatively straightforward method to developing a stress period setup to enter as much variation for river levels as possible. The stage data is updated online daily and this represents the maximum resolution possible for stress periods with respect to stage elevation. The temporal discretisation scheme best suited will depend on the questions asked of the floodplain model. For example, if a long-term effect on the floodplain is investigated then monthly mean river levels are likely to be adequate. If simulating drought conditions where there is very little stage fluctuation over significant periods of time then a yearly mean approach to the stress period setup is also adequate. If the short term effects of an environmental watering event or a flood is the aim, then an adaptive scheme may be warranted.

Spatial discretisation may have to factor into the choice of temporal scheme as well. One reason for this is the inclusion of solute transport, which places stability constraints on time steps particularly when mass conservative solution techniques are implemented in advective flux calculations. The 125 m by 125 m horizontal discretisation used in Goyder Floodplain Model is well suited to the flow velocities that occur within layer 2. It is currently unclear how a change in horizontal discretisation will affect cumulative salt accessions as the near-river salinity currently plays a major role in the concentration of water entering the river.

If freshwater lenses were investigated then the horizontal spatial discretisation currently implemented will undoubtedly have to be refined. This ultimately results in longer model run times as the number of nodes increases and with it the number of time steps necessary to maintain stability.

The current vertical discretisation of 20 m reduces variation in the solute distribution beneath the river and the vertical gradients. In this model the river and reservoir cells are linked to nodes in layer 2 which places the entry point for water from the river effectively 12 metres below surface. This has a significant effect on the salinity distribution adjacent the river in both the vertical and horizontal directions.

Very little change is observed in the solute distribution in layer 2 during the 20 year simulations. This raises questions as to the significance of the initial solute distributions effect on the cumulative salt load to river. The initial concentrations were derived using a model with a static river level over a 1000 year period, which is adequate for a synthetic model. When developing a real world floodplain model field evidence of variation in solute distribution will be mandatory if a relatively accurate estimate of salinity impact can be attributed to any one event.

Very little separates the reservoir cells performance from the river cell. Both packages function similarly and have virtually identical input requirements. The reservoir cell approach does offer some benefits during model development in the form of reduced user input but the lack of support for solute transport simulation in MT3DMS makes it a poor replacement for the river cell in this instance.

4.4.2 Scenario 2- Evapotranspiration Scenario

Scenario 2, the evapotranspiration scenarios, covered seven scenarios as discussed in Section 4.3.2, specifically:

- Evapotranspiration varying yearly (Scenario 2A)
- Evapotranspiration varying monthly (Scenario 2B)
- Evapotranspiration varying monthly with a 0.5 m reduction in extinction depth (Scenario 2C)
- Evapotranspiration varying monthly with a 0.5 m increase in extinction depth (Scenario 2D)
- Spatial varying evapotranspiration rate including monthly variation (Scenario 2E)
- Spatial varying evapotranspiration rate and extinction depth, including monthly variation (Scenario 2F)
- Evapotranspiration varying monthly, using a non-linear evapotranspiration versus depth function, ETS curve (Scenario 2G)

The snap shots in time provided in the following sections were taken at December 1996 and December 2006 of the following:

- Water balance
- Watertable
- Salinity in the surface aquifer
- Actual evapotranspiration lost
- Flux of water into and out of the river

4.4.2.1 Water Balance, Salt Balance

The dominant components of the water balance (Figure 8.34 and Figure 4-35) for all scenarios are evapotranspiration out of the floodplain and inflow through the general head boundaries (see p91 for comments on the labeling of the water balance figures). Evapotranspiration outflow is usually larger than general head boundary inflow, however for Case C without a lock, the inflow through the general head boundaries is slightly larger than evapotranspiration. This is due to Case C having a gaining river, and the additional inflow from the general head boundaries is translated into flow from the floodplain to the river.

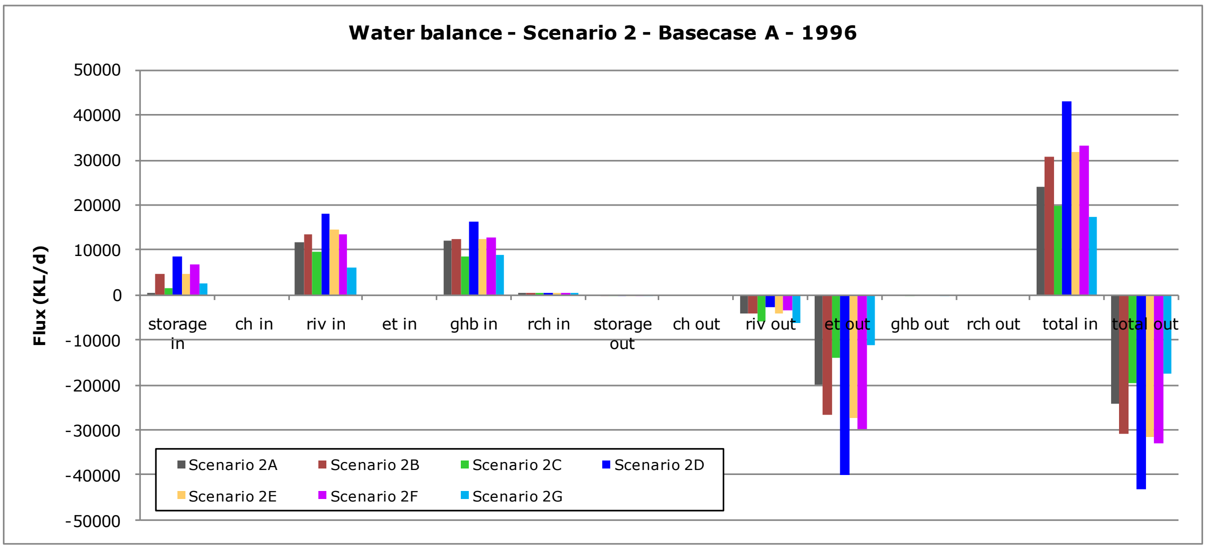


Figure 4-34 Water Balance at 1996 for Case A with a lock

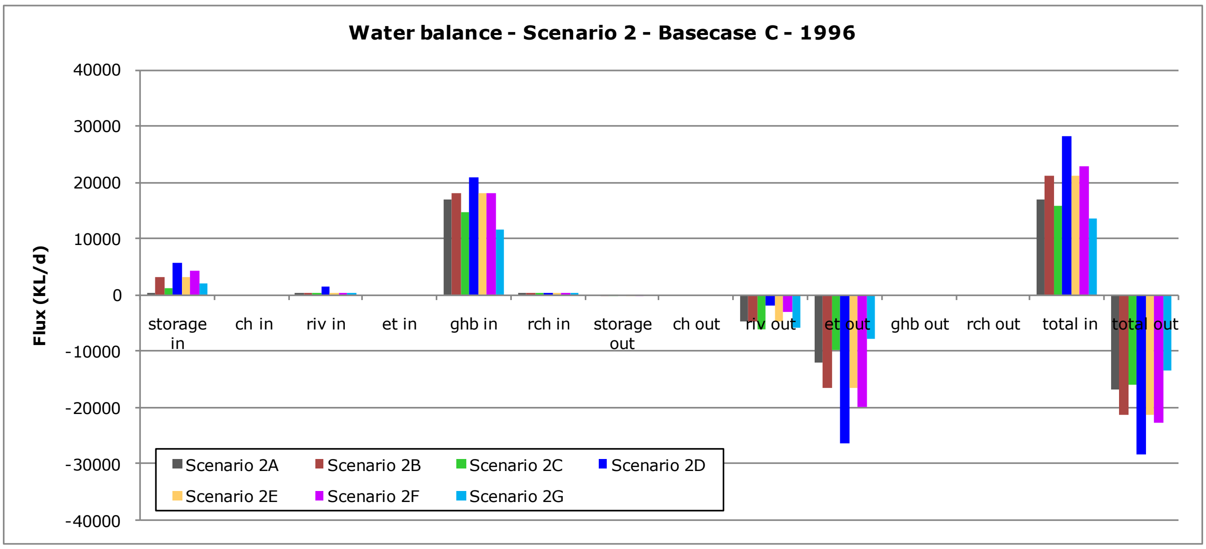


Figure 4-35 Water Balance at 1996 for Case C without a lock

By comparing between the scenarios (Scenario 2A to 2G) the following observations are made:

- Increasing the temporal resolution (yearly to monthly) increases the volume of inflows and outflows, particularly evapotranspiration
- Manipulating the extinction depth of the evapotranspiration had a direct and significant impact on the water balance
  - This was more pronounced in models without the lock
  - Models with a lock have areas near the lock where watertables are held high due to flow around the lock. These higher water levels remain accessible to evapotranspiration even under reduced extinction depth. Hence, changing extinction depth has less of an impact on models with a lock.
- There was minimal difference between the spatially-homogeneous rate and the spatially-varied evapotranspiration rate (scenario 2B compared to scenario 2E).
- Combining the spatial variation of evapotranspiration rate with spatially-varying extinction depth increased the evapotranspiration compared to both the non-spatially varied scenario and the evapotranspiration rate varied scenario (scenario 2F compared to scenario 2B and 2E).

- This is consistent with the observation that changing the extinction depth has a significant impact on the water balance
- Changing the evapotranspiration function (scenario 2G) showed the greatest reduction in evapotranspiration (compared to scenario 2B) out of all the runs in the evapotranspiration scenario, except for Case B without a lock.

Since the evapotranspiration function in scenario 2G used the same extinction depth as scenario 2B, the critical factor that influences evapotranspiration is not the extinction depth by itself. If a comparison is made between the two evapotranspiration curves used in these scenarios (Figure 4-36), then it can be seen that at depths below 0.05 m, the normal linear ET curve will extract more evapotranspiration than the ETS curve used in scenario 2G.

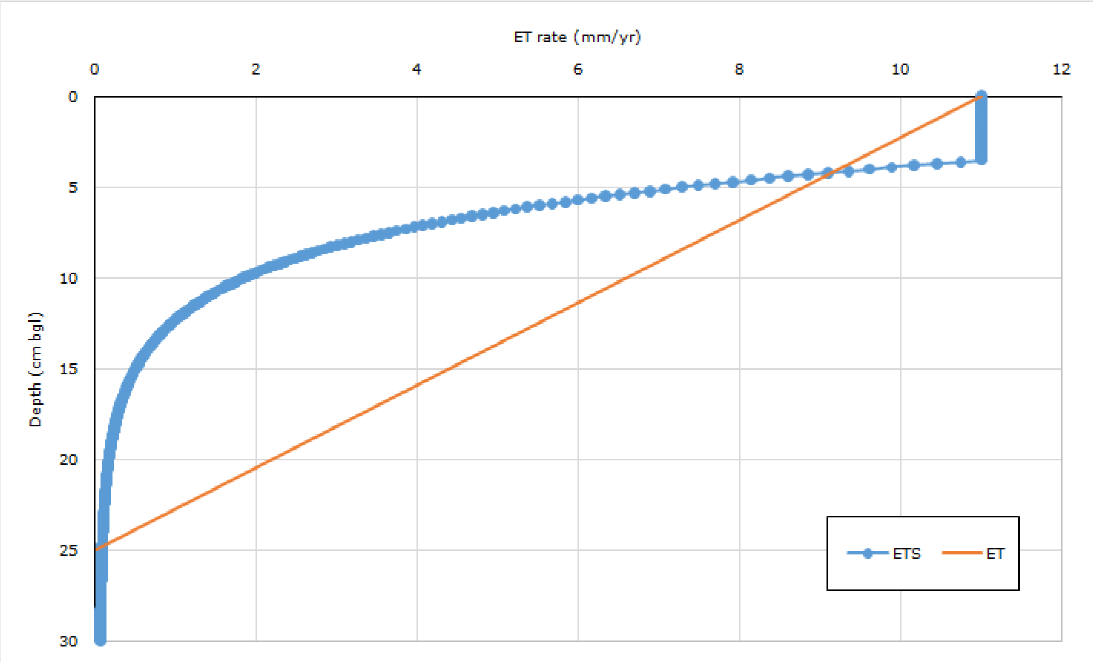


Figure 4-36 Comparison between linear ET curve and ETS curve with theoretic water table

4.4.2.2 River condition and salt load

Changes in evapotranspiration have the capacity to change how the surface water system interacts with the groundwater (Figure 4-37 and Figure 4-38). For all cases with the lock in place, the river is a losing river under scenarios 2A (yearly-varying evapotranspiration) and 2B (monthly-varying evapotranspiration). Increasing the extinction depth (scenario 2D), spatially varying the evapotranspiration rate only (scenario 2E) and then the extinction depth as well (scenario 2F), all show an increase in the losing condition of the river. This result is consistent with the hydrographs which showed that these scenarios either lowered the groundwater levels or made minimal change (scenario 2E) to them. Models where the extinction depth was reduced (scenario 2C) or the relationship between evapotranspiration rate and depth was non-linear (scenario 2G) showed a move toward a gaining river, for most if not all of the time.

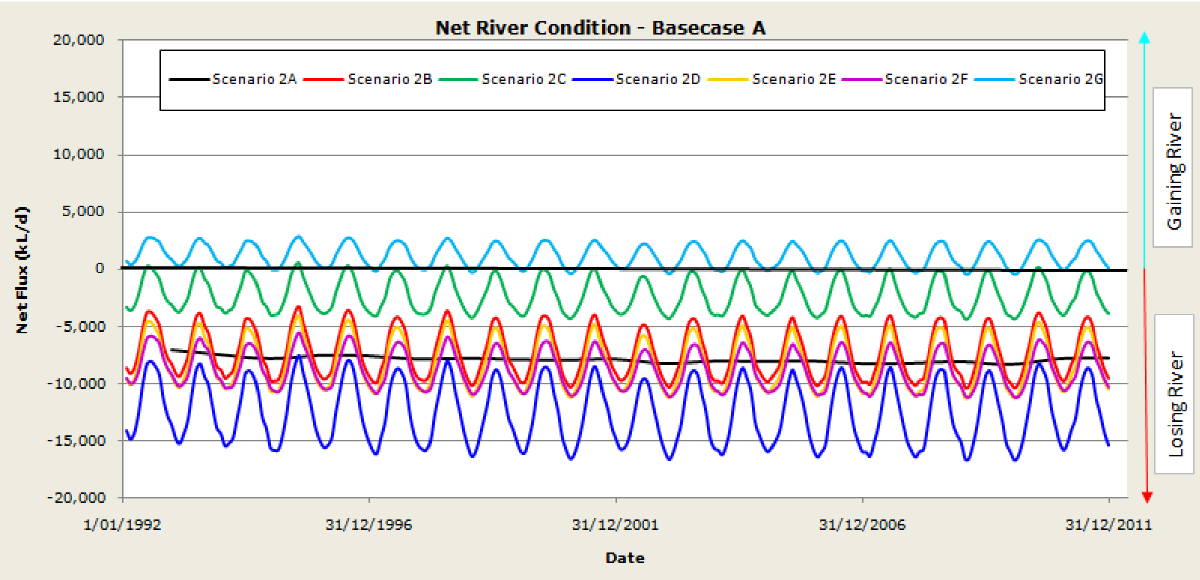


Figure 4-37 River Condition at 1996 for Case A with a lock

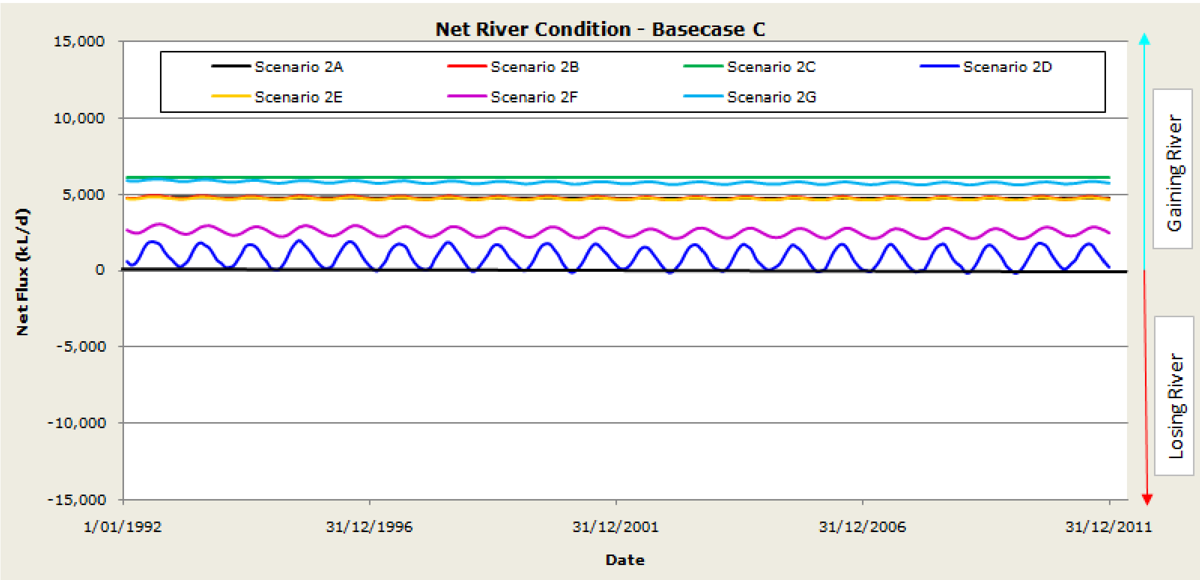


Figure 4-38 River Condition at 1996 for Case C without a lock

Most models with monthly-varying fluctuations show a seasonal fluctuation in the relationship between the river and the groundwater. The only exceptions to this is seen in Case C without a lock, with monthly variation (scenario 2B) and a reduction in extinction depth (scenario 2C). This is also the only scenario where using the non-linear ETS curve (scenario 2G) is below (less gaining) the monthly varying (using a linear ETS curve) evapotranspiration (scenario 2B). This is likely caused by a combination of the reduced floodplain width to the north of the river reducing the area over which evapotranspiration is effective and the differences between the two evapotranspiration curves at the top of the curve. Figure 4-36 (above) shows a comparison of the two curves: when the watertable is shallow the non-linear curve is able to extract more evapotranspiration. This then lowers the watertable to a point where evapotranspiration is less effective, allowing the water table to recover, resulting in a fluctuating groundwater level. The fluctuating groundwater level causes changes in the relationship between the river and the groundwater. For the scenarios where the evapotranspiration rate is too low to impact on the watertable, no fluctuation occurs and hence no fluctuation in the relationship between the river and the groundwater.

For all cases, reducing the extinction depth (scenario 2C) and using a non-linear relationship between evapotranspiration rate and depth (scenario 2G) have significantly greater cumulative salt load to the river compared to other scenarios. Unlike the Scenario 1, time discretisation appears to have minimal effect on the cumulative salt load to the river, with little difference between the yearly varying (scenario 2A) and the monthly variation (scenario 2B) of evapotranspiration rate.

Cumulative salt flux comparisons, between the solute transport model and the cumulative salt load calculated based on an assumed constant salinity, generally show that the two approaches yield the same trends (Figure 4-39 to Figure 4-42). For models without locks, the cumulative salt loads calculated by two methods are also similar, however for the models with locks, the solute transport results generally show about a third less salt entering the river. This is likely due to the flow around the lock, which would be predominately fresh, which would not be calculated as fresh in the assumed salinity calculations.

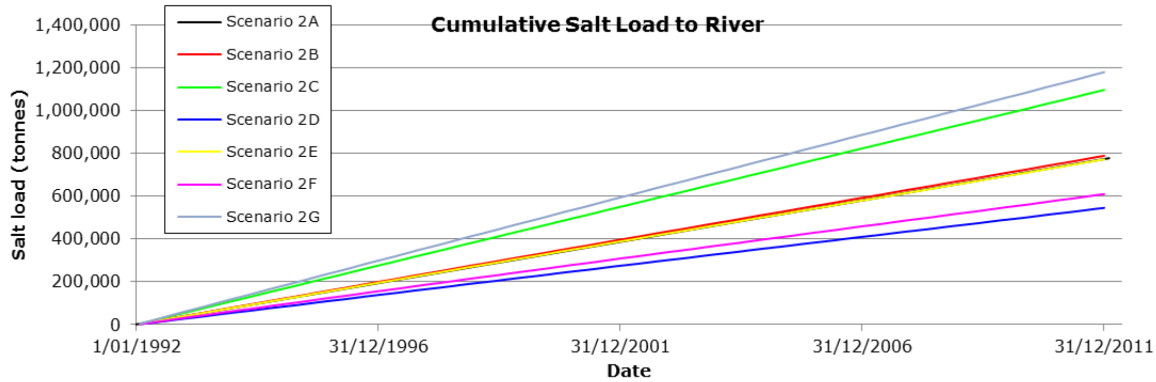


Figure 4-39 Cumulative Salt Load for Case A with a lock, based on flow model

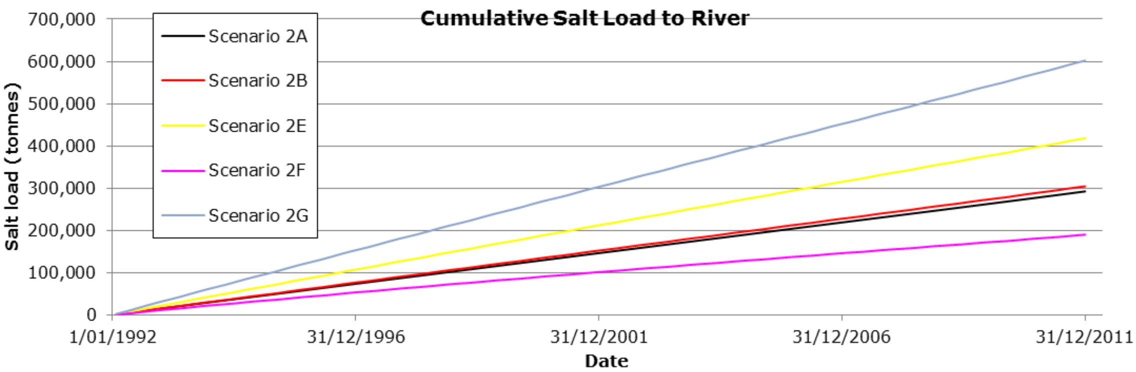


Figure 4-40 Cumulative Salt Load for Case A with a lock, based on solute model

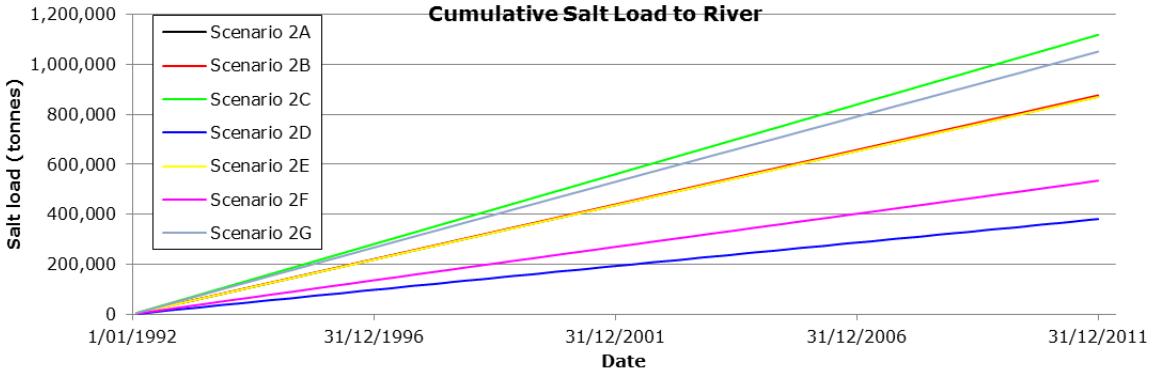


Figure 4-41 Cumulative Salt Load for Case C without a lock, based on flow model

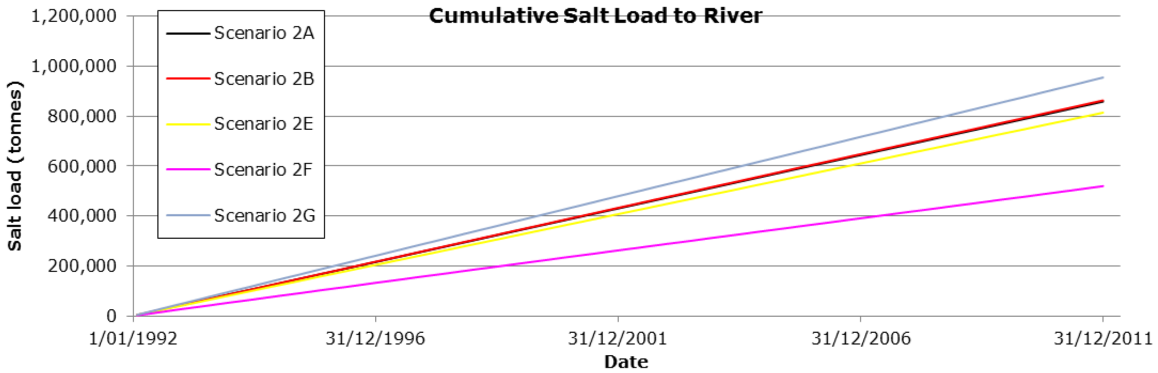


Figure 4-42 Cumulative Salt Load for Case C without a lock, based on solute model

4.4.2.3 Hydrographs

Hydrographs (Appendix C) show the following:

- Simulations with monthly time steps show more fluctuation in the groundwater levels than the yearly simulations
- Reducing the extinction depth raises the groundwater level, conversely increasing the extinction depth reduces the groundwater level
- Spatially varying the evapotranspiration rate only had little impact in the groundwater level
    - The exception to this was Case C with a lock, where the spatial variation reduced the groundwater level in the highland area
- Spatially-varying the extinction rate and depth generally reduced the groundwater level, often to levels similar to groundwater levels in the increased extinction depth scenario (scenario 2D). The response to this scenario was variable between bores, dependent on the changes to evapotranspiration at the bores location.

Using the ETS function (scenario 2G) resulted in a general rise in groundwater levels. This was less evident further away from the river in models without locks as well as downstream of the locks in models with locks. The ETS function also resulted in a greater amplitude of fluctuation in groundwater levels near the river, particularly in the Coonambidgal Formation (Layer 1); the exception to this was Case C without a lock, where groundwater levels north of the river were mostly unchanging over time, similar to scenario 2B. This is reasonable for the groundwater levels in the regional aquifer, where it is expected that the depth to water is greater than the extinction depth, however it was also seen in the bore closest to the river. It is likely that this is a result of the evapotranspiration being insufficient on the small area of floodplain between the highland area and the floodplain to impact on the groundwater level. This is consistent with the results shown in the above section which indicate that the net river flux shows little variation over time in these scenarios.

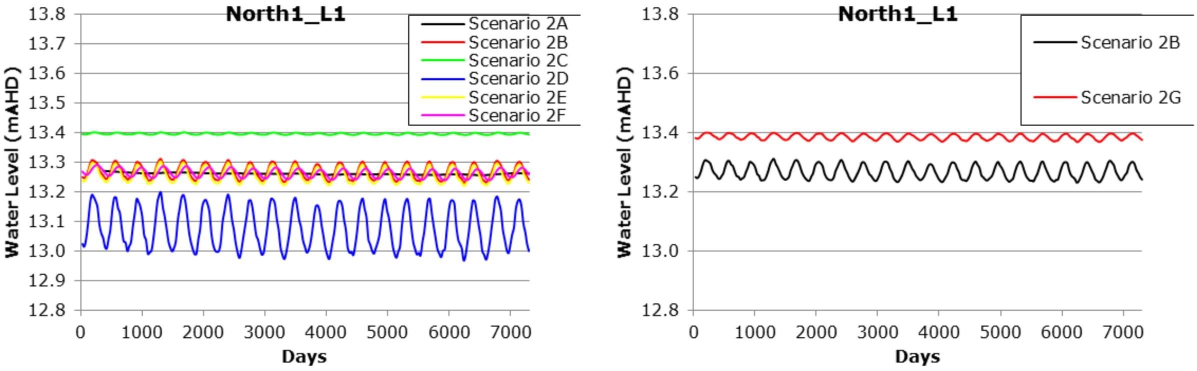


Figure 4-43 Hydrographs for Case A with a Lock

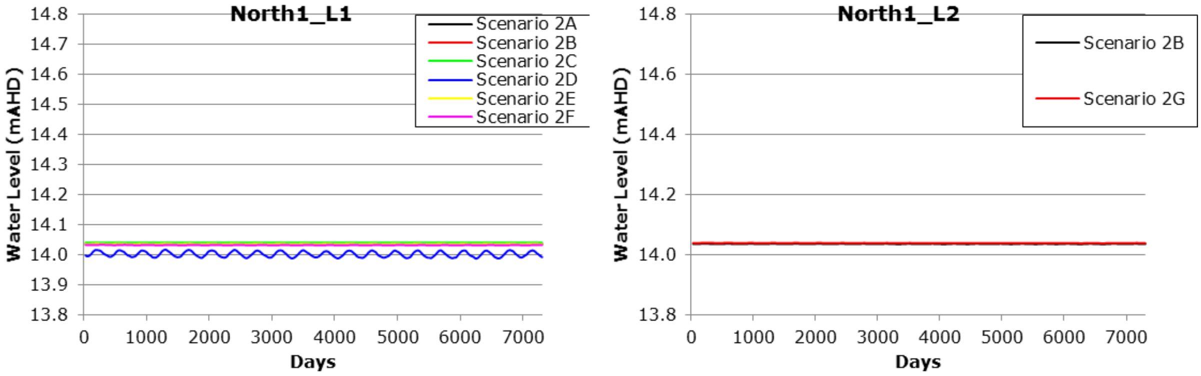


Figure 4-44 Hydrographs for Case C without a Lock

4.4.3 Scenario 3 - Inundation Scenario

Scenario 3, the inundation scenarios, covered three scenarios as discussed in Section 4.3.3, specifically:

- Inundation of the floodplain and wetland with a constant rate (Scenario 3A)
- Inundation of the wetland only with a constant rate (Scenario 3B)
- Inundation of the floodplain and wetland with a varying rate (Scenario 3C)

For comparison, some of the inundation results have been compared to the time-varying river levels (monthly) without inundation (Scenario 1A).

The snap shots in time provided in the following sections were taken at December 1996 and December 2006 of the following:

- Water balance
- Watertable
- Salinity in the surface aquifer
- Actual evapotranspiration lost
- Flux of water into and out of the river

4.4.3.1 Water Balance, Salt Balance

The dominant components of the water balance (Figure 4-45 to Figure 4-48) for all inundation scenarios are:

- Inflow from the river (river into the model – riv in)
- Inflow from the general head boundaries (larger component in Case B and C); and
- Outflow from evapotranspiration on the floodplain (et out).

See p91 for further explanation of the labels used in water balances figures.

Under wet conditions, models without a lock where floodplain inundation has been applied (Scenario 3A and 3C), recharge also becomes a dominant water balance component, greater than the inflow from the river or general head boundaries. For models with a lock, these components remain dominant regardless of whether the floodplain is inundated or not. This is driven by the relationship between the river level and the topography. For models without a lock, inundation occurs along the length of the river to the height of the river level. For the models with a lock, only the area that is topographically less than the downstream river level will be inundated. This means that the models without a lock are inundated over a greater area and thus have larger recharge.

Under dry conditions, the inflow from the general head boundaries is a significantly greater component of the water balance than the inflow to the floodplain from the river. As the river condition changes between cases, the significance of the general head inflow changes. Case A (losing river) has general head inflows approximately three times larger than the inflow to the floodplain from the river. Case C (gaining river) has general head inflows approximately eight times larger than the inflow to the floodplain from the river.

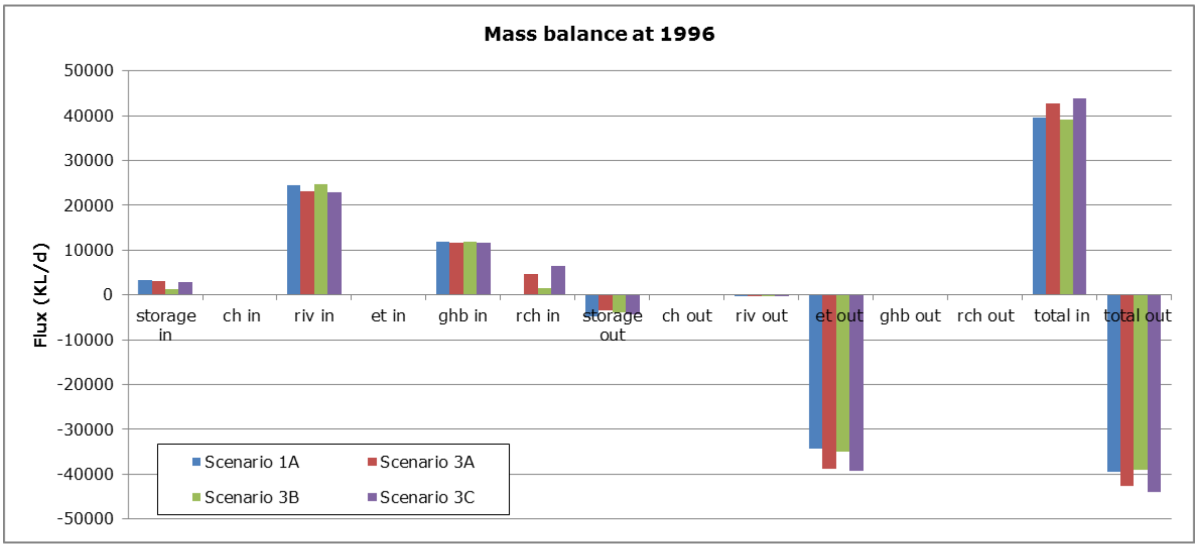


Figure 4-45 Water balance for Case A with lock under wet conditions.

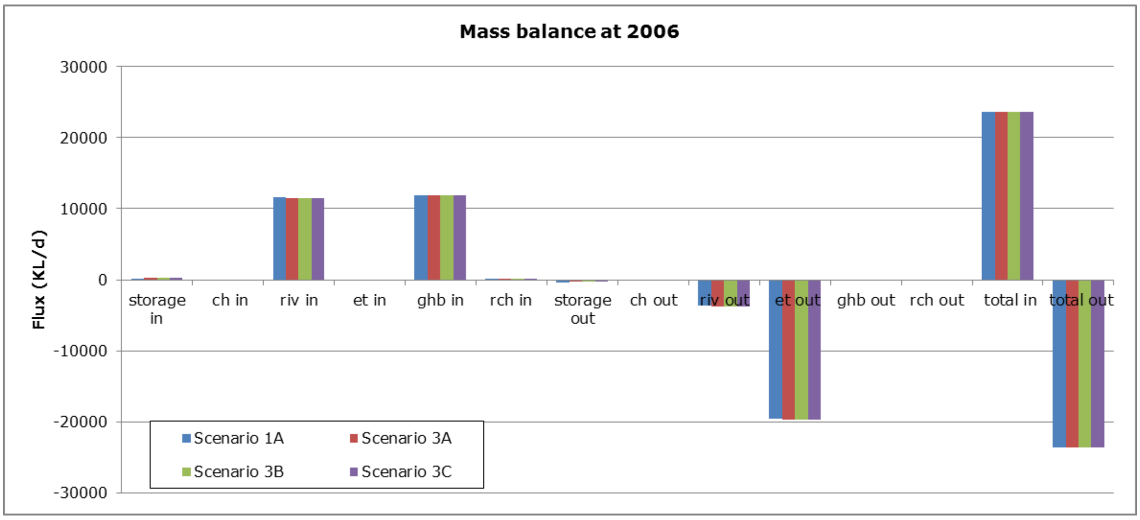


Figure 4-46 Water balance for Case A with lock under dry conditions

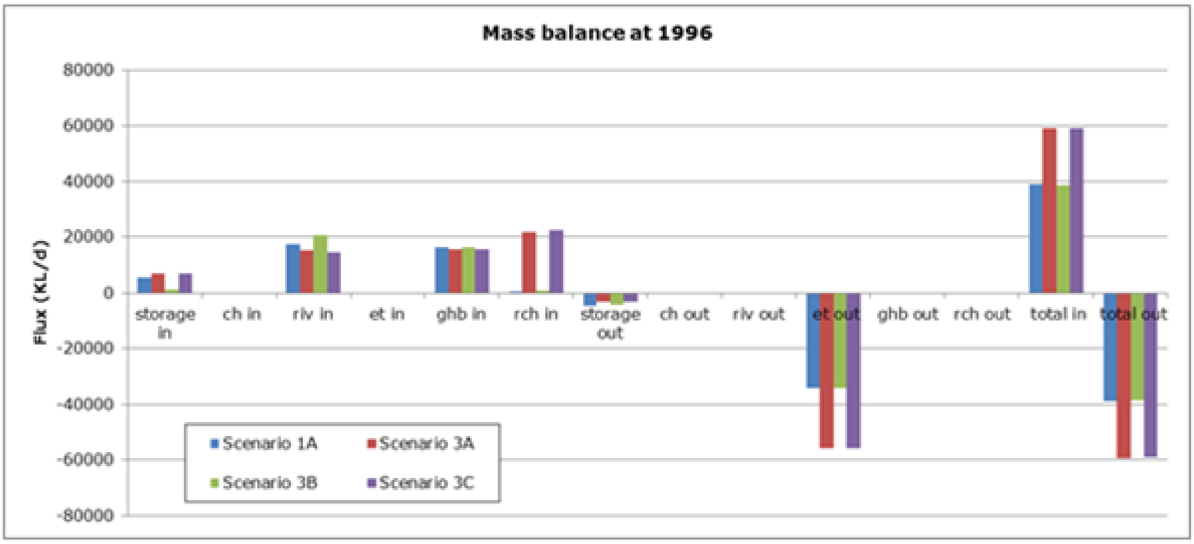


Figure 4-47 Water balance for Case C without lock under wet conditions

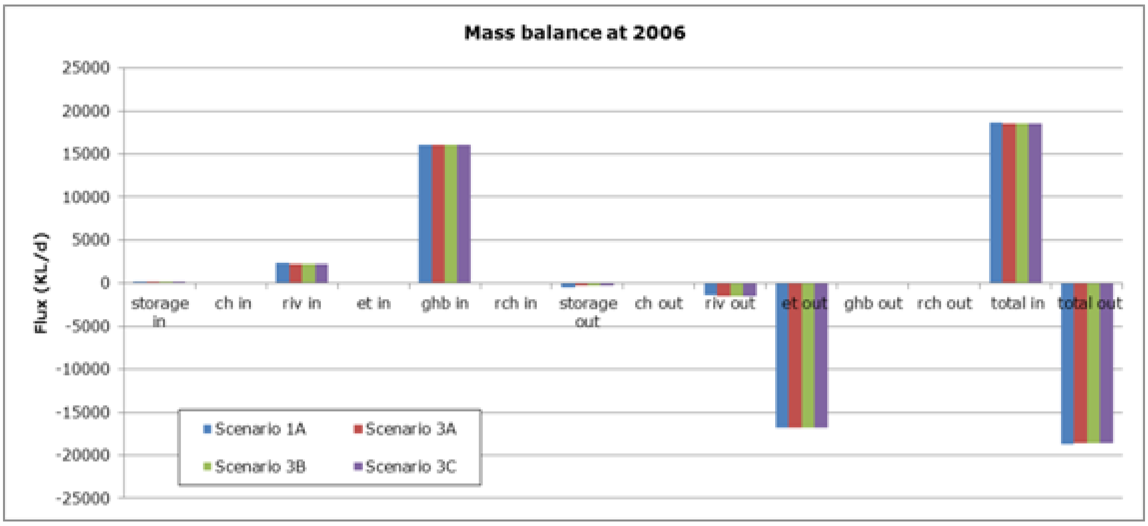


Figure 4-48 Water balance for Case C without lock under dry conditions

4.4.3.2 River condition and salt load

Inundation does not change the condition of the river over the long term, however, on a short term basis, inundation impacts the gaining river conditions that occur post flood. This can be seen in the net river condition figures (Figure 4-49 and Figure 4-50) which show increased flux into the river post flood compared to river fluctuation only (scenario 1A). While this is seen in both the models with and without a lock, the impact is greater in the model without a lock, as can be seen in the cumulative salt load to river (Figure 4-51 and Figure 4-52).

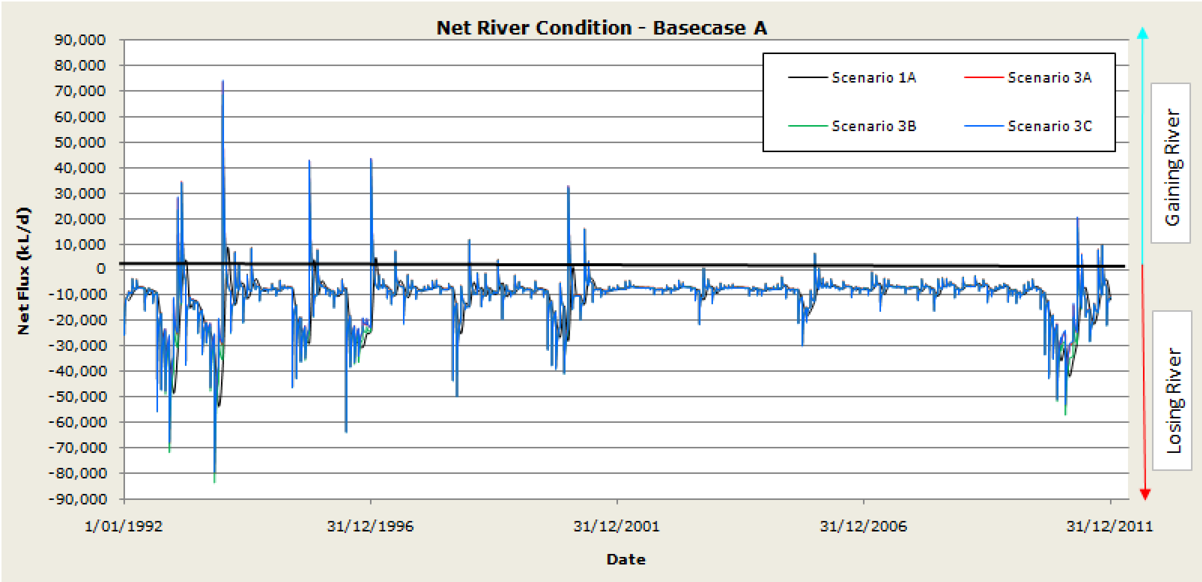


Figure 4-49 River Condition for Case A with Lock

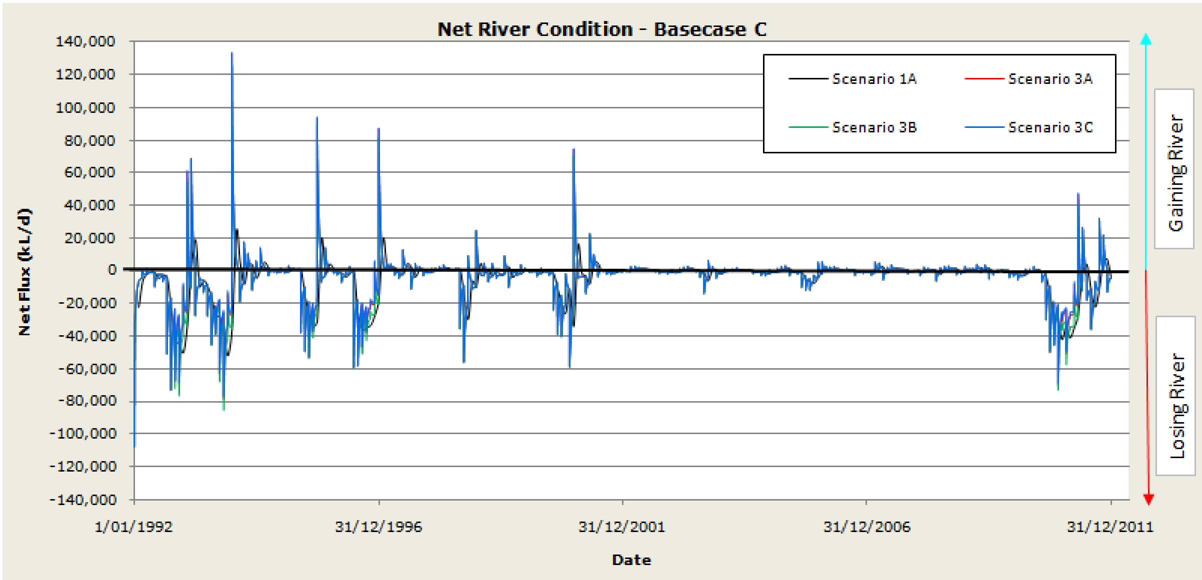


Figure 4-50 River Condition for Case C without Lock

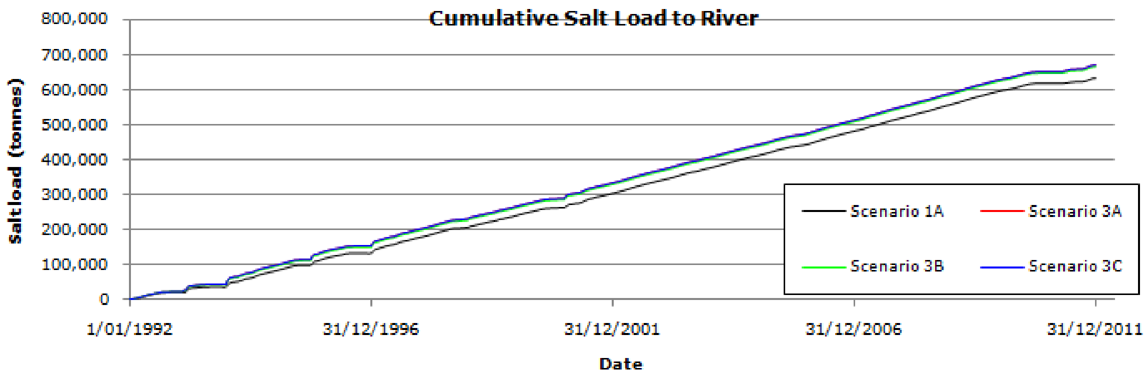


Figure 4-51 Cumulative Salt Load for Case A with a lock, based on flow model

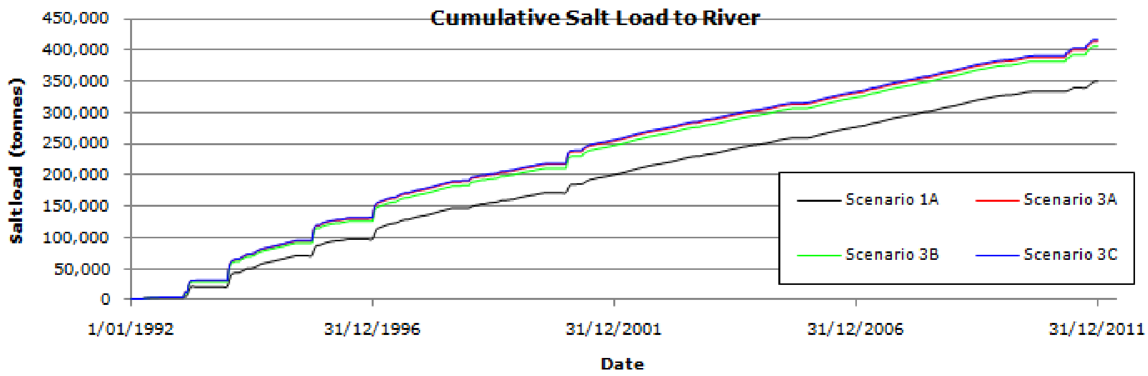


Figure 4-52 Cumulative Salt Load for Case C without a lock, based on flow model

The only time the river is able to gain salt is under gaining river conditions. By looking at the difference between models with and without inundation (scenario 3 and scenario 1A), the impact of the floodplain inundation on salt load to river becomes evident. While the total salt load is greater in the model with the lock, the impact of inundation is greater in the model without the lock. Case A and B almost double the cumulative salt load to the river over a 20 year period under inundation, and Case C experiences a 20% increase. It is reasonable to expect Case C to have less of an increase as it has a large salt load to river prior to adding inundation, due to its design as a gaining river.

Cumulative salt flux comparisons between the solute transport model and the cumulative salt load calculated based on an assumed constant salinity and the flow model, show that the two approaches yield the same general trends, however the solute transport models estimate salt loads which are consistently less than the salt loads based on the constant-salinity assumptions. In the models with the lock, the solute transport cumulative salt load was usually less than half the cumulative salt load based on the constant-salinity assumption over a 20 year period. The difference is less for the models without a lock, indicating that the freshwater input to the system from inundation is an important component of the overall groundwater input to the river. This clearly demonstrates the importance of solute transport in a dynamic floodplain environment.

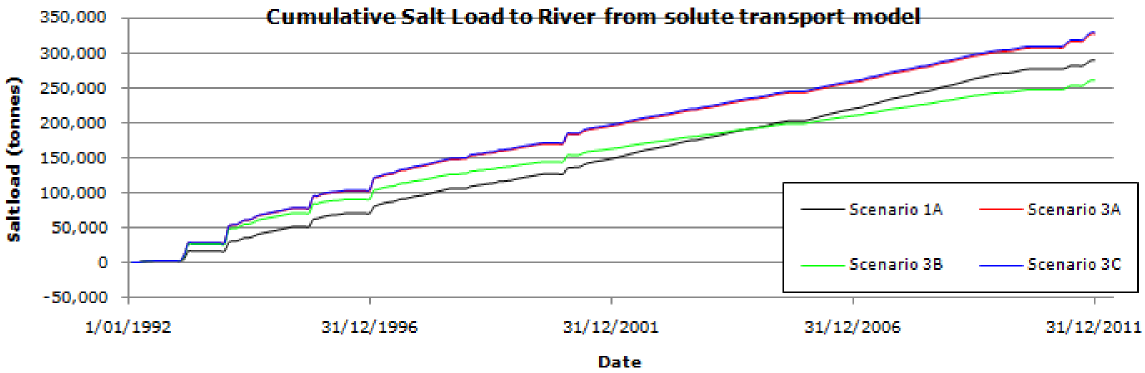


Figure 4-53 Cumulative Salt Load for Case A with a lock, based on solute model

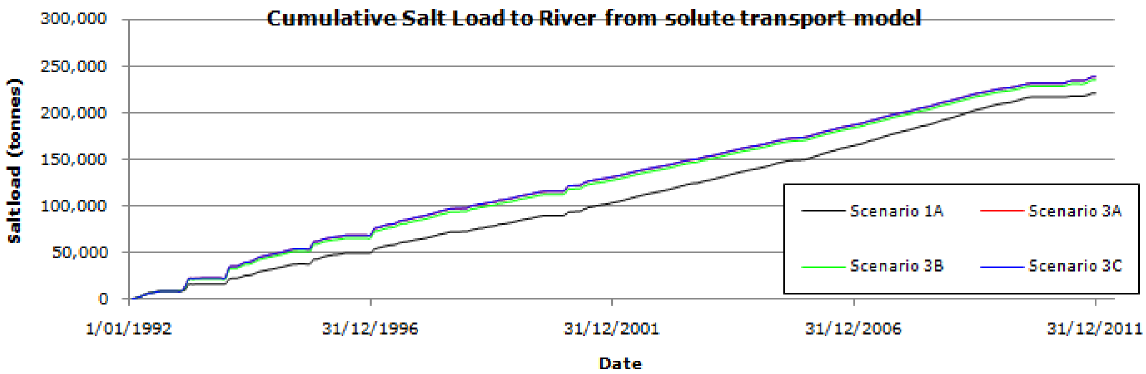


Figure 4-54 Cumulative Salt Load for Case C without a lock, based on solute model

4.4.3.3 Hydrographs

Hydrographs (Appendix C in Woods, 2015b) for all cases show increased watertable levels during increased flow and inundation. Increases were greater away from the river, with the exception of Case C where the northern highland area showed no noticeable change in water level. Floodplain inundation had greatest impact on increasing watertable level, while the impact of wetland inundation was not evident in observation bores, with the exception of the South 2 bore in the models with a lock. This bore is located at the northern edge of the wetland and was the only bore to show a difference in water levels between river level fluctuations only (scenario 1A) and wetland inundation (scenario 3B). This may indicate a highly-localised impact of wetland inundation.

Vertical groundwater gradients from the Monoman Formation (layer 2) to the Coonambidgial Formation (Layer 1) were reduced near the river under floodplain inundation. However this gradient increased further away from the river. For models with a lock this was only the case downstream of the lock.

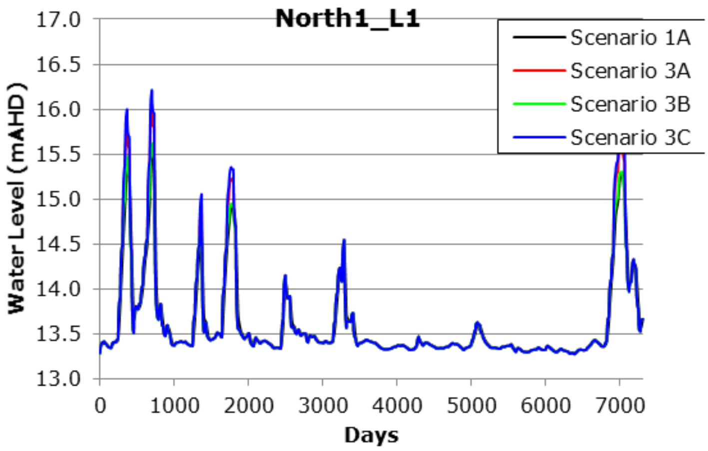


Figure 4-55 Hydrographs for Case A with a Lock – near river, downstream of Lock

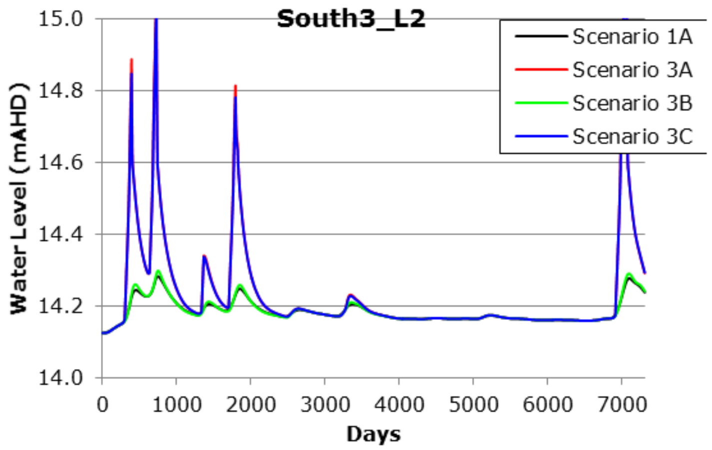


Figure 4-56 Hydrographs for Case C without a Lock – approximately 1 km from the river

4.4.3.4 Wet and dry conditions

Figure 4-45 to Figure 4-48 show key water balance components during a wet period and a dry period for the three inundation setups investigated.

Under wet conditions (December 1996) the models show that the river is predominately losing, less so upstream of the lock for models with a lock. Potentiometric heads are generally higher than under dry conditions. For models without a lock, increases are most evident near the river, which corresponds to the area under inundation for these models. For models with a lock, increases are most evident downstream of the lock and around the wetland, which corresponds to the area under inundation for these models. Heads are lower under wetland-only inundation (scenario 3B) compared to floodplain and wetland inundation (scenarios 3A and 3C). Evapotranspiration is up to 3 mm/day (50 m^3/day) concentrated on the areas near or being inundated, i.e. near the river in models without a lock, or downstream of the lock and around the wetland in models with a lock.

Under dry conditions (December 2006) the models show the following: For models without a lock, the river is predominately neutral (Case A) or gaining (Case B and C). For models with a lock the river is losing above the lock and gaining below the lock. Groundwater heads are generally lower than under wet conditions, particularly downstream of the lock, for models with a lock. Evapotranspiration is up to 0.6 mm/day (10 m^3/day) concentrated on a narrower near river floodplain and wetland area, compared to under wet conditions. Evapotranspiration is up to 1.5 mm/day (24 m^3/day) upstream of the lock (for models with a lock), similar to the same area under wet conditions. Note all contour and flooded contour plots of have the x and y axis in meters with the units and descriptions for the flooded regions given as the plot title.

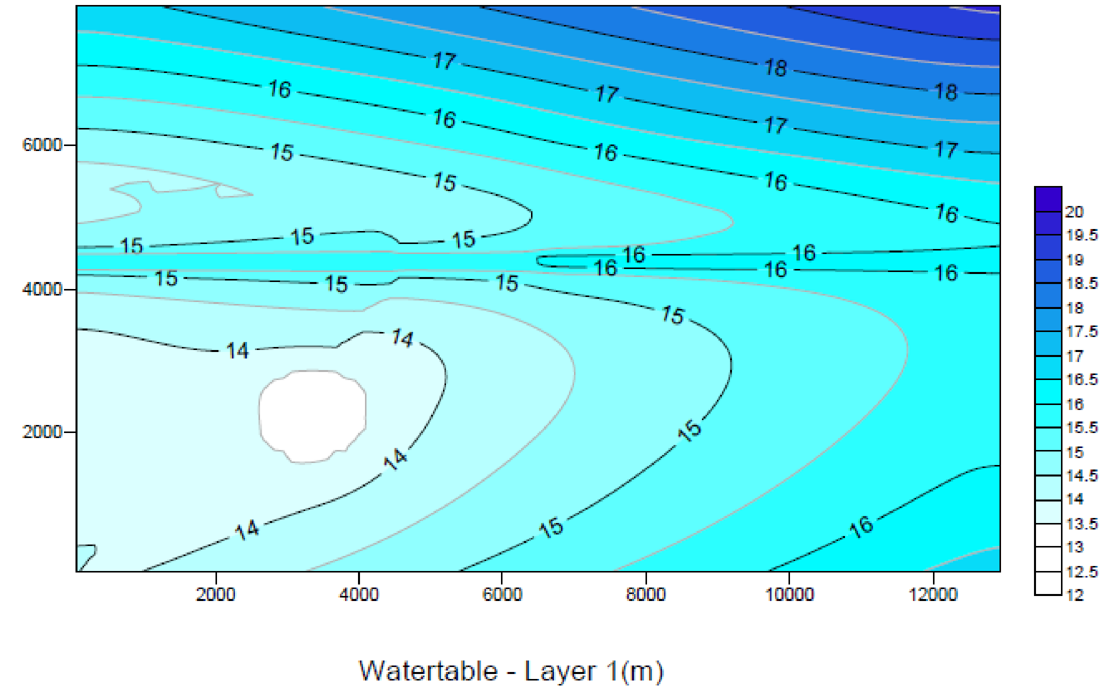


Figure 4-57 Watertable for Case A with a Lock at December 1996, Scenario 3A.

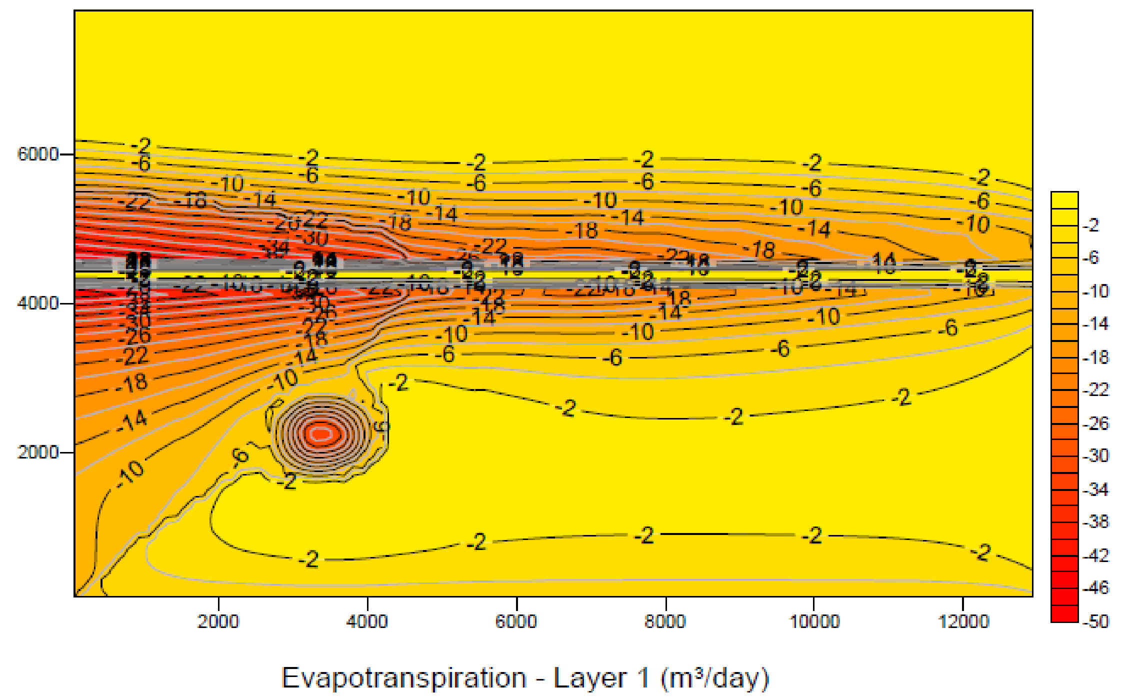


Figure 4-58 Evapotranspiration for Case A with a Lock at December 1996, Scenario 3A

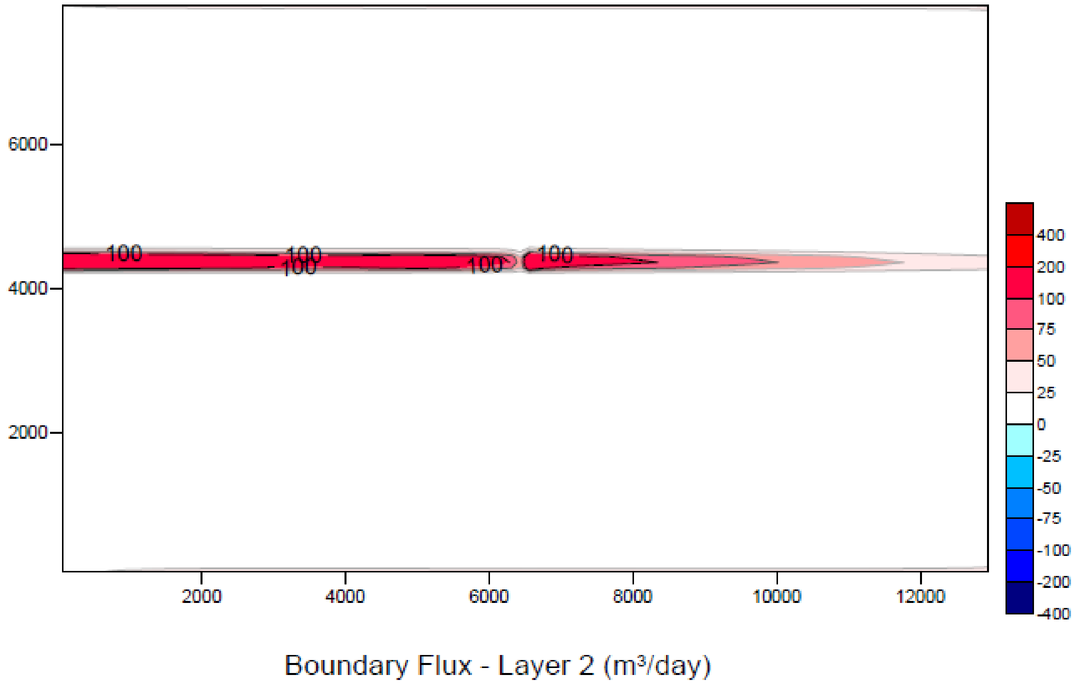


Figure 4-59 River boundary flux for Case A with a Lock at December 1996, Scenario 3A

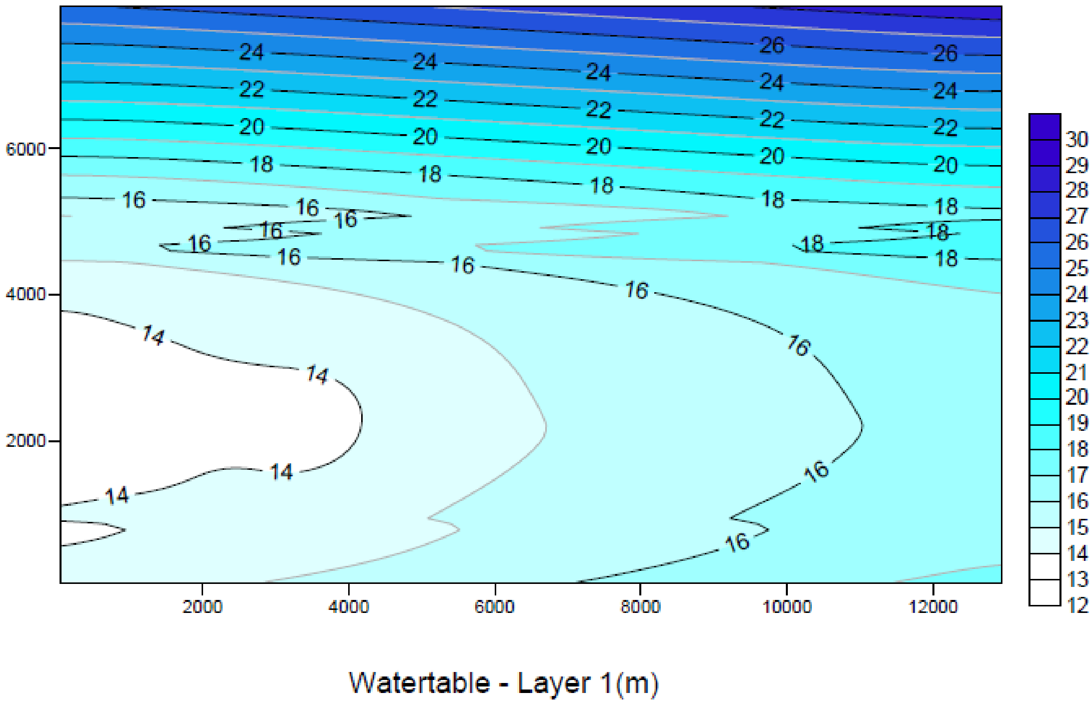


Figure 4-60 Watertable for Case C with a Lock at December 1996, Scenario 3A

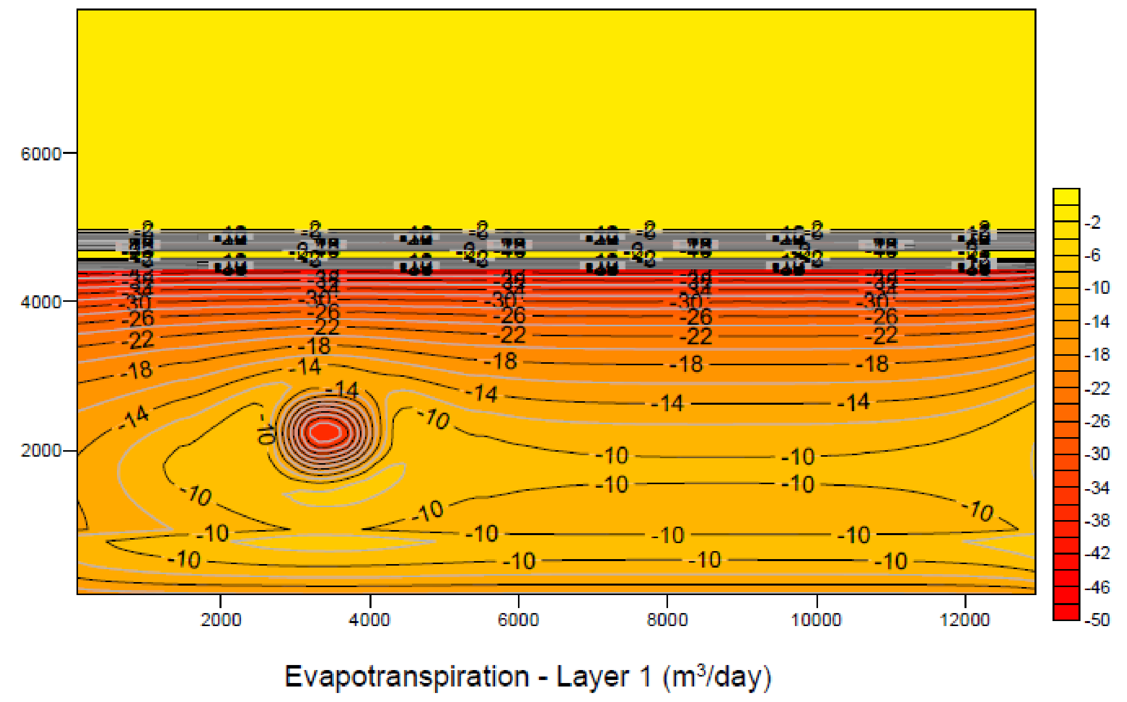


Figure 4-61 Evapotranspiration for Case C with a Lock at December 1996, Scenario 3A

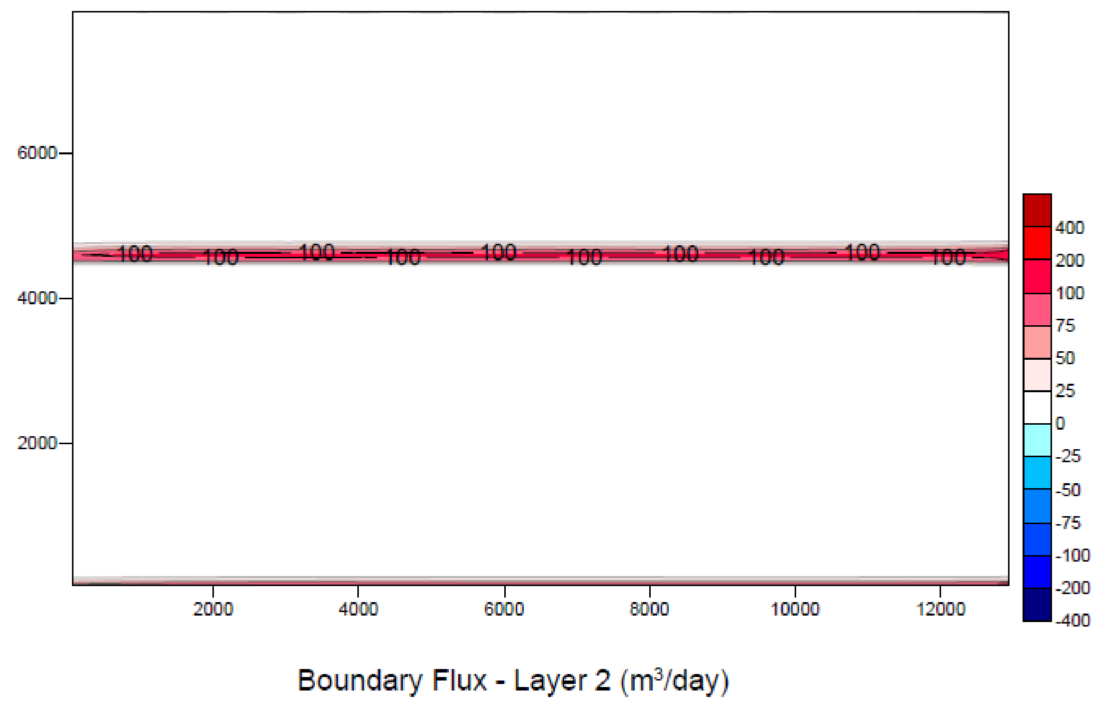


Figure 4-62 River boundary flux for Case C with a Lock at December 1996, Scenario 3A

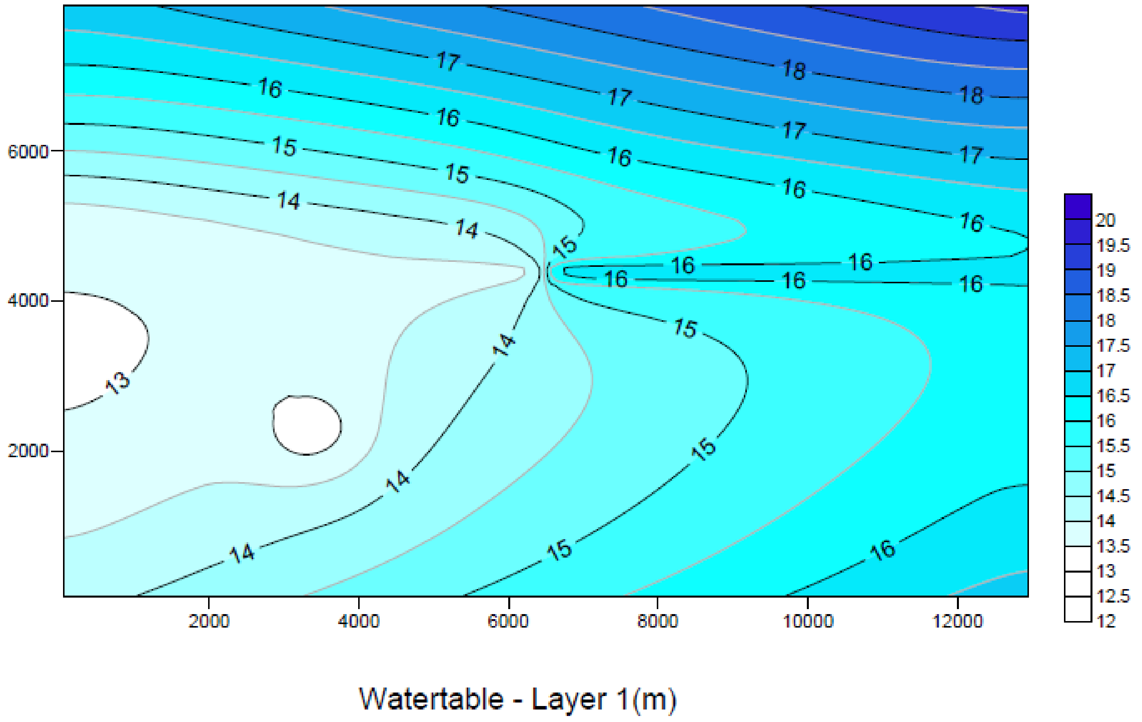


Figure 4-63 Watertable for Case A with a Lock at December 2006, Scenario 3A

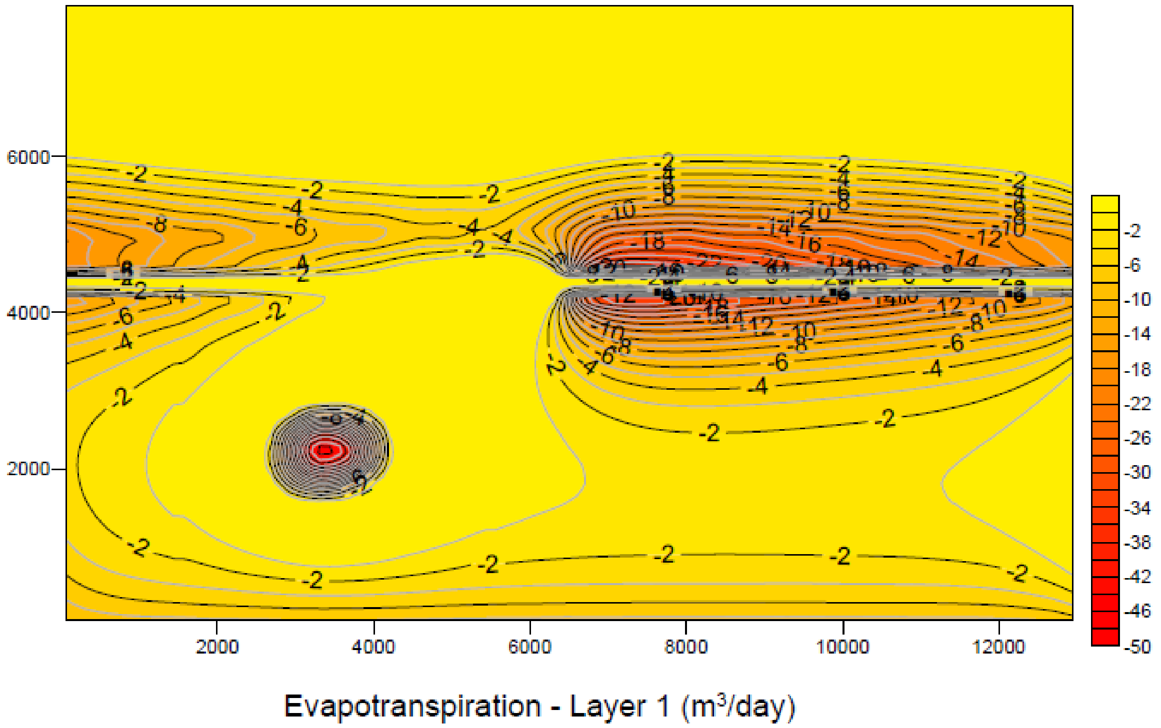


Figure 4-64 Evapotranspiration for Case A with a Lock at December 2006, Scenario 3A

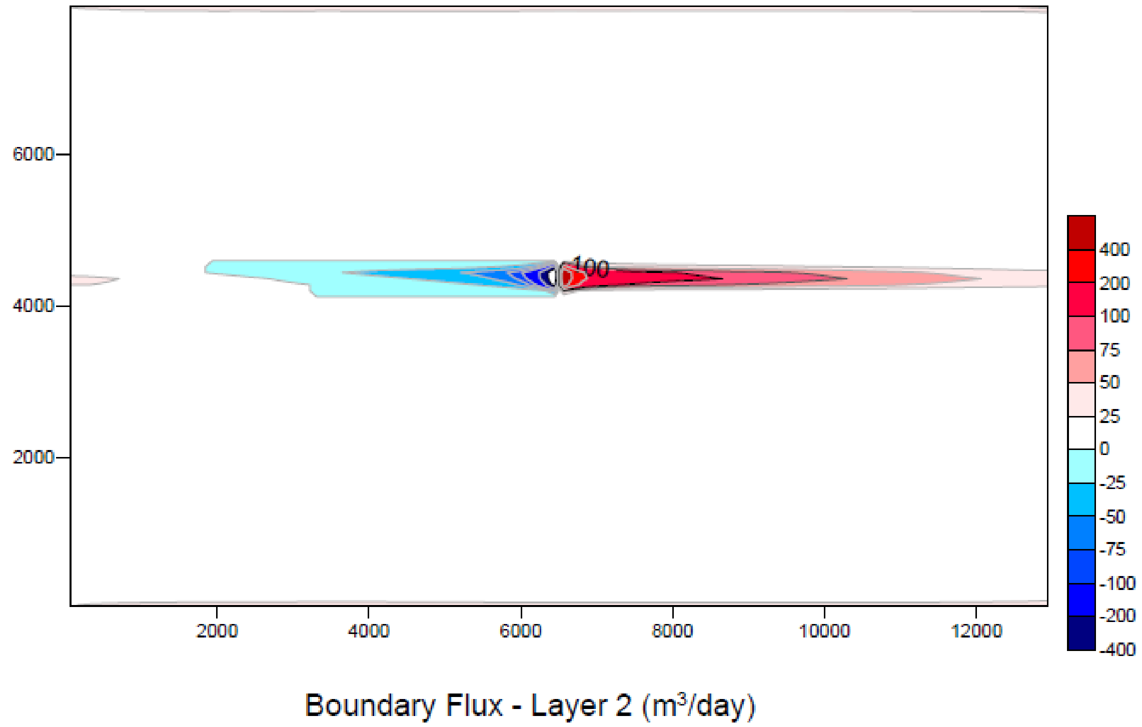


Figure 4-65 River boundary flux for Case A with a Lock at December 2006, Scenario 3A

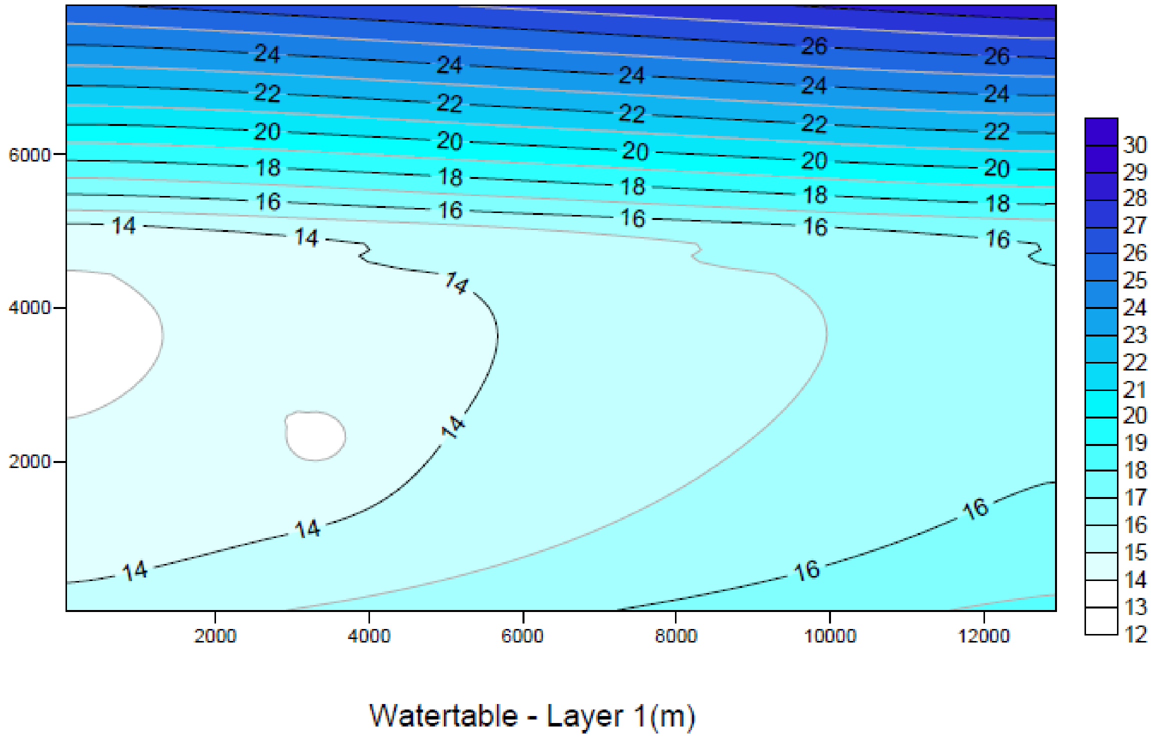


Figure 4-66 Watertable for Case C with a Lock at December 2006, Scenario 3A

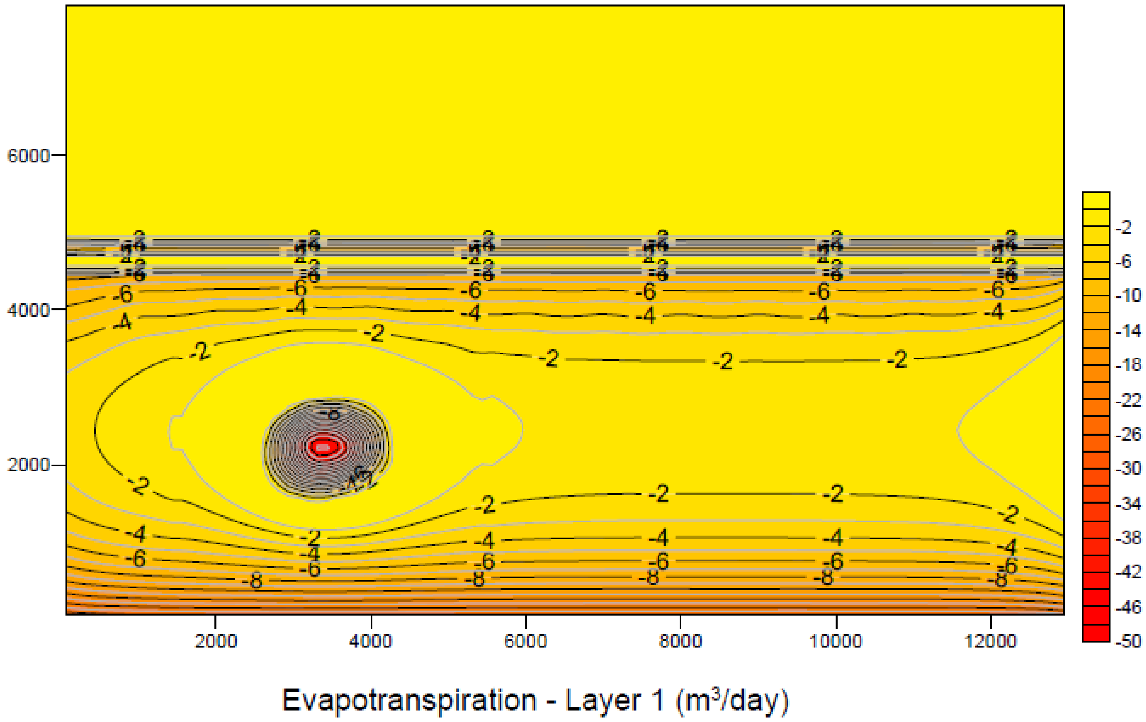


Figure 4-67 Evapotranspiration for Case C with a Lock at December 2006, Scenario 3A

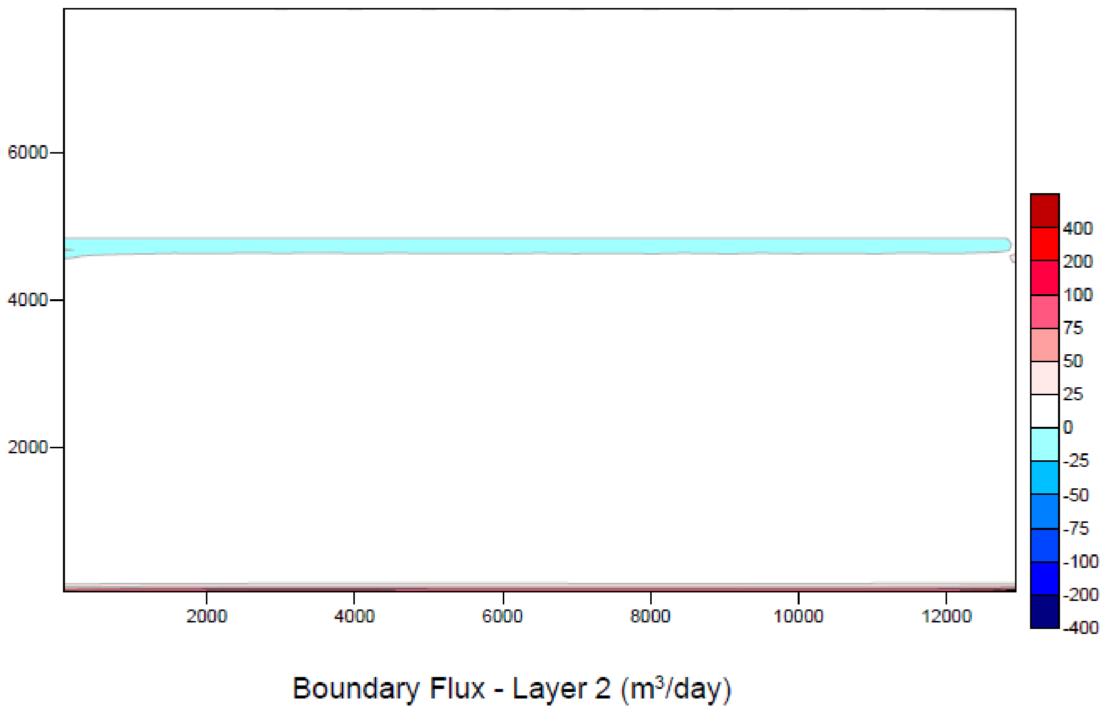


Figure 4-68 River boundary flux for Case C with a Lock at December 2006, Scenario 3A

4.5 Discussion

Based on the simulations presented in this chapter, we recommend the following when developing a MODFLOW model of the salinity dynamics in the study area:

- Use MODFLOW2005 with the NWT solver and UPW package, as this simulates the wetting and drying of the Coonambidgal Formation more consistently than earlier versions of MODFLOW. MODFLOW-Surfact may also be used, but is not compatible with the most-commonly used solute transport code for MODFLOW, MT3DMS.
- Simulate solute transport explicitly, rather than estimating salt transport in groundwater by multiplying flux by a constant salinity.
- Use monthly or sub-monthly stress periods which are shorter during floods, in order to better capture the behaviour. Appendix A of Woods (2015b) includes a Python code for selecting stress periods based on river level.
- Vary the maximum ET on at least a monthly basis, to capture seasonal changes.
- Develop an understanding of how evapotranspiration relates to depth and choose an appropriate representation of this, which may include using a piecewise linear ET function.
- Carefully consider the representation of the surface topography: in a floodplain, minor variations in elevation have a large impact on ET.
- Consider the use of the MODFLOW ET package recently developed by Doble *et al.* (2015), which uses a lookup table for recharge with respect to land use, climate, soil, vegetation type, and depth to water, based on unsaturated zone modelling.
- Identify areas of likely higher inundation recharge by examining AEM, vegetation and soil data.
- Incorporate outcomes of work done to address conceptual gaps, e.g. revised inundation recharge rates from unsaturated zone investigations, and constraining the model to match AET observations

Further investigations could be made with the generic models presented here, for example:

- Further analyse the results of the simulations conducted so far
- Conduct sensitivity analyses to key parameters, including but not limited to:
    - Murray Group transmissivity, simulating the full saturated thickness of the Murray Group, rather than one aquifer unit
    - Solute transport parameters such as longitudinal and transverse dispersivity and porosity
    - Initial conditions for the salinity
- Simulate the circumstance where vertical upwards flow is the predominant source of regional groundwater flux into the floodplain aquifer, e.g. at Woolpunda, by increasing the potentiometric head in layer 4 and increasing the vertical hydraulic conductivity of the regional aquitard in layer 3
- Consider the use of the 1D unsaturated zone simulation package for MODFLOW as an alternative to representing inundation as specified recharge
- Trial adaptive stress periods for ET, similar to what has been investigated for the representation of changing river levels. Consider how to select stress periods which are based on both changing river levels and ET simultaneously.
- Trial the use of the stream package for the River Murray, considering whether this a better representation that would allow easier linking to surface water models such as Source

- Trial a simulation in which the River Murray curves through the landscape
- Further investigate the importance of numerical discretisation on model results using versions of MODFLOW which allow for local grid refinement, e.g. MODFLOW-USG and iMOD
    - Horizontal discretisation (area of cells)
    - Vertical discretisation of the Monoman Sands aquifer for solute transport simulation, to represent variations in salinity with aquifer depth
    - Disctretisation around a lock
- Trial using GSFLOW, USGS's coupled surface water and groundwater flow model based on MODFLOW

The following model tasks are noted as useful but are out of scope of the present study:

- Evaluation of codes and methods that could be used to support a Salinity Risk Framework
- Evaluation of codes and methods as rapid assessment tools for salinity impacts
- Comparison of MODFLOW model of a floodplain with a HydroGeoSphere or other fully-integrated model

5 Surface water model development

Peter Cook, Chris Turnadge, Robert Bridgart, Trevor Pickett and David Rassam

5.1 Introduction

Hydrologic river models are frequently used to examine how changes in river management will impact on flows and salt loads. However, these river models do not include many important groundwater processes, or they include them only in highly simplified manner (Chapter 3). Thus, while there is a need to improve the simulation of salt dynamics in the River Murray floodplain in groundwater models, there is also a need to improve the way that river models simulate this process. It is acknowledged that predictions from simple hydrologic river models are likely to be less accurate than predictions from 2D hydrodynamic river-floodplain models or numerical groundwater models. However, it is clear that these simple river models have an important part in water management, and will continue to be used.

There is an ongoing debate in the scientific literature concerning the roles of simple and complex models. Sometimes simple models contain the same processes as complex models, but do not allow for the same degree of spatial variation in parameters. Often though, simple models do not include all of the processes that are included in complex models. However, if simple models are chosen, we need to ensure that:

1) The simple models include the important processes
2) Processes in the simple model are included sufficiently accurately

Availability of data should not be a consideration in the choice between simple and complex models. If a process is important, then it should be included in the model, even though we may not have data for all of the relevant parameters (Doherty & Simmons, 2013). It is better to assume a value for a model parameter (and then to carry out sensitivity analysis of model predictions to this assumed value), than to ignore a process completely. However, it is also sometimes a useful strategy to begin with simple models, and progress to more complex models only as required. The inability of a simple model to reproduce field data might be a justification for developing a more complex model (Hill, 2006). This is one of the key advantages of hydrologic river models (Chapter 3).

This aim of this activity was to develop a simple model of salt transport from a floodplain to a river through a groundwater system. There are a number of processes which operate within river floodplain environments, and some of these are discussed in Woods (2015a), and displayed diagrammatically on Figure 5-1. The focus of the current chapter is only on the pathway indicated in red on Figure 5-1 – the groundwater flow path from the wetland to the river. Although we also consider regional groundwater flow to the river, we do not consider possible flow from the river to the wetland, or bank storage exchange. Ideally, the simple model developed in the chapter would be able to be included in river hydrologic models that are used for water management purposes. The inclusion of floodplain salt transport routines in a regional river flow model would therefore allow the interaction between river flow, and flow regulation and salt loads to be examined. Of course, a number of other processes are important for water management, but we do not specifically consider them here. Although the approaches considered in this chapter are general and in principle could be incorporated into a number of different models, in this chapter we used the Source model (Welsh *et al.*, 2012) as a platform for testing the approach.

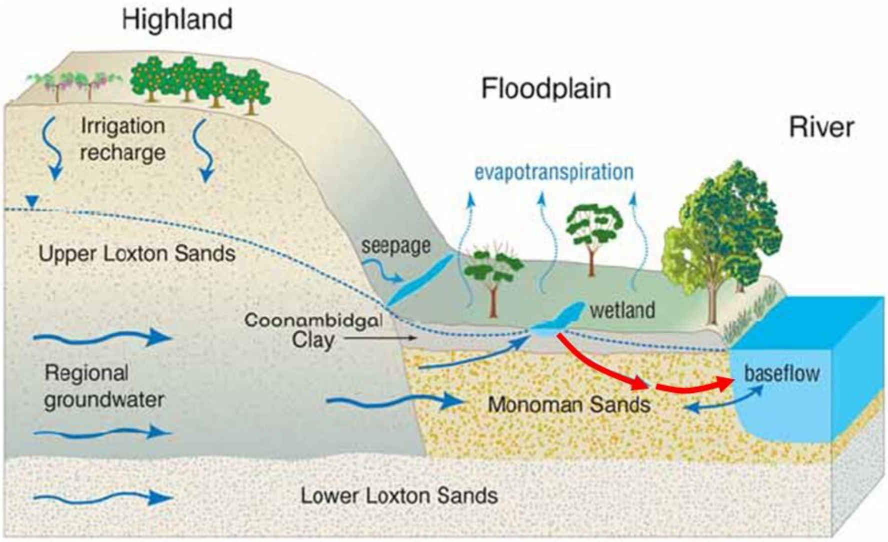


Figure 5-1 Conceptual model of the River Murray floodplain (Source: Holland *et al.*, 2005). The red arrow indicates the flow and transport pathway which is the focus of this chapter.

5.2 Theory

5.2.1 Introduction

Consider a pipe initially filled with saline water. Freshwater enters the pipe at the one end. Almost immediately, saline water is pushed out of the other end of the pipe. Sometime later, freshwater emerges out of the pipe.

The same thing happens when infiltration into an aquifer occurs close to a river. Firstly, the increase in pressure caused by the infiltrating water causes an increase in groundwater flow to the river. This groundwater flow is water that was already present in the aquifer, analogous to the saline water in the previous example. Sometime later, the infiltrating water discharges to the river (analogous to the fresh water described above). For convenience, we refer to the first process as the *pressure* response, and the second as the *transport* response.

Although these two processes are intimately linked, it is sometimes convenient to consider them separately. It is also convenient to conceptualise the river-wetland-groundwater system in terms of a number of possible end-member configurations. Thus the wetland could represent a point or it might be long and parallel to the river, and the leakage rate beneath the wetland might be small or large relative to the regional groundwater flow rate. How we would represent the interaction between the wetland and the river is slightly different for each of these cases.

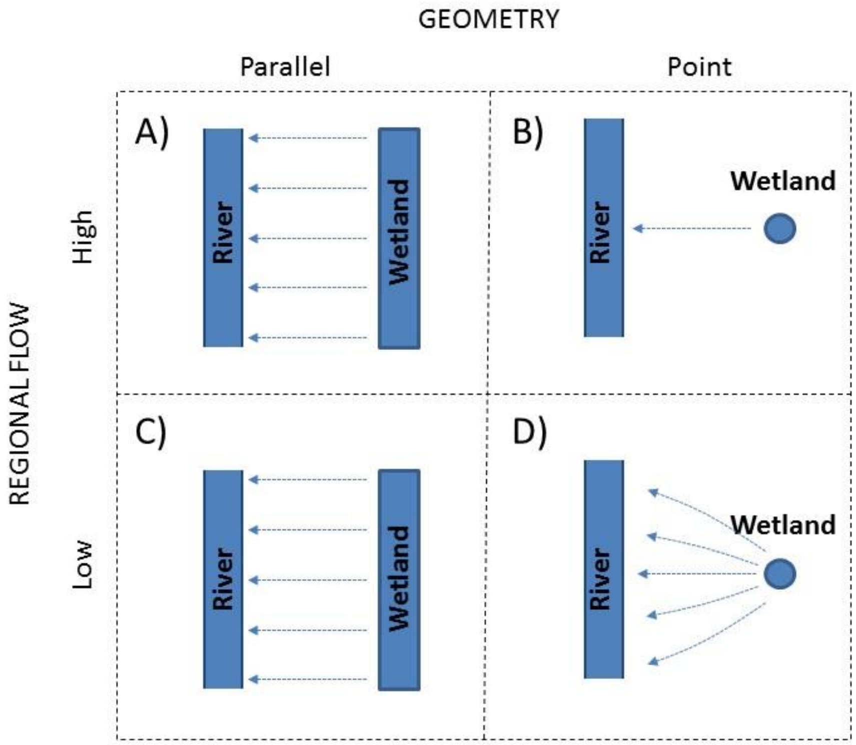


Figure 5-2 Conceptual diagram of different possible configurations of the river-wetland-groundwater system, in which regional groundwater flow is directed towards the river. Cases A and C represent long, thin wetlands that approximately parallel the river, while cases B and D represent smaller, more discrete wetlands. Broken lines depict approximate groundwater flow lines, depending on whether regional groundwater flow rates are high (Cases A and B) or low (Cases C and D) relative to wetland leakage rates.

5.2.2 Pressure Response

The change in groundwater discharge rate to a river due to the pressure effect created by a short pulse of recharge some distance from the river can be expressed:

$$\Delta q_r(t) = Q \frac{x}{2t\sqrt{\pi D t}} \exp\left(\frac{-x^2}{4Dt}\right) \tag{5-1}$$

where $\Delta q_r(t)$ is the change in discharge to the river (m^3/day), Q is the volume of recharge (m^3), x is the distance of the recharge source (i.e., wetland) from the river (m), D is the aquifer hydraulic diffusivity (m^2/day) and t is time since the recharge pulse (Knight *et al.*, 2005). (Note that $D = K\,b/S$, where K is the aquifer hydraulic conductivity, b is the aquifer thickness and S is the specific yield.) More frequently, the cumulative response function is used:

$$\Delta Q_r(t) = Q \operatorname{erfc}\left(\sqrt{\frac{x^2}{4Dt}}\right) \tag{5-2}$$

where $\Delta Q_r(t)$ is the cumulative change in discharge to the river (m^3) (Glover and Balmer, 1954; Knight *et al.*, 2005). Equation 5-2 also gives the instantaneous change in river discharge for the case of a continual recharge pulse (i.e., a step increase in recharge). Equations 5-1 and 5-2 are plotted in Figure 5-3. In this case (x = 100 m, D = 10 m^2/day), the peak response occurs after 167 days. Equations 1 and 2 approximately describe the change in flux with time, irrespective of whether the wetland represents a discrete point or a long river-parallel feature. Of course, Equations 5-1 and 5-22 assume that the change in recharge occurs at a fixed distance x metres from the river. Other equations have been developed where the change in recharge occurs over a larger area (Cook *et al.*, 2003; Knight *et al.*, 2005).

The salt flux to the river arising from this recharge pulse can then be calculated as either:

$$\Delta\, m_r(t) = Qc_g \frac{x}{2t\sqrt{\pi Dt}} \exp\left(\frac{-x^2}{4Dt}\right) \tag{5-3}$$

or:

$$\Delta M_r(t) = Q\; c_g\; \mathrm{erfc}\left(\sqrt{\frac{x^2}{4Dt}}\right) \tag{5-4}$$

where $m_r(t)$ and $M_r(t)$ are the instantaneous and cumulative salt fluxes to the river (units of g/day and g, respectively), and are determined from the water flux (Equations 5-1 and 5-2) and the salt concentration in the groundwater (c_g).

Equations 5-1 to 5-4 can be used to determine the change in water flow and salt mass to the river whether recharge is positive (e.g., an increase in infiltration or recharge) or negative (e.g., a decrease in recharge or increase in groundwater pumping). They have been most widely used for calculating the effect of groundwater pumping on rivers, but have also been used for estimating the change in salt load to the river from changes in dryland recharge (Cook & Connor, 2002; Cook *et al.*, 2003; Wang *et al.*, 2005) and from recharge from irrigation developments (Miles *et al.*, 2001; Knight *et al.*, 2005).

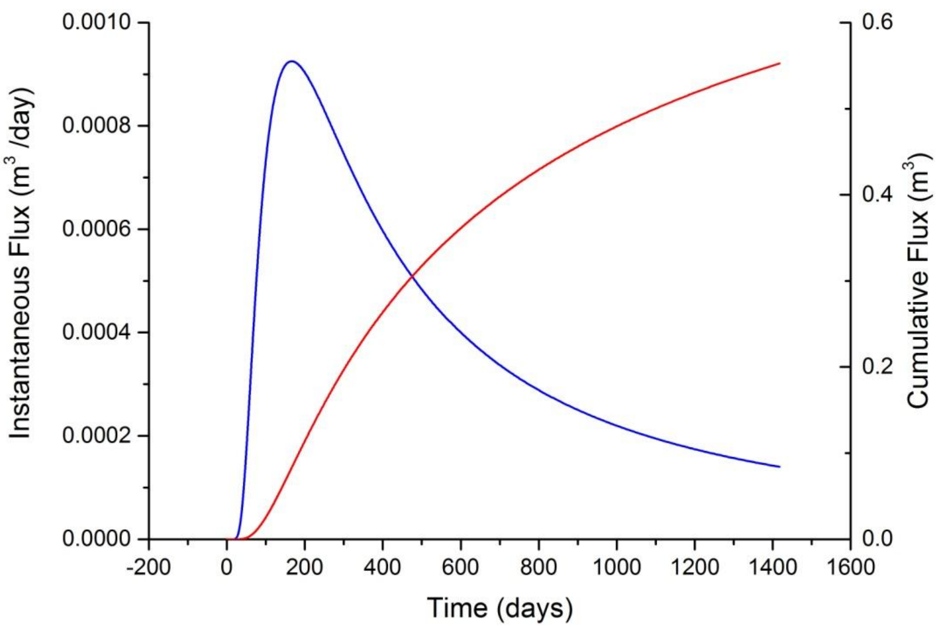


Figure 5-3 Change in flow of water to the river following a short recharge pulse of 1 m^3 at a distance of 100 m from the river (D= 10 m^2/day). Both instantaneous (Equation 5-1; blue line) and cumulative (Equation 5-2; red line) fluxes are shown.

The problem with using this approach for determining the change in salt load to the river from an increase in recharge is that the concentration of recharge is not explicitly considered. There is only a single term for concentration in these equations and this represents the concentration of ambient groundwater. Equations 5-3 and 5-4 thus implicitly assume that the concentration of recharge is the same as the ambient groundwater concentration. If this is the case, then the mass balance will be correct (i.e., the total salt mass to the river will be equal to the total salt mass in recharge). Because the travel time for water from the point of recharge to the river will be longer than the pressure response time, the assumption that the concentration of recharge is equal to the ambient groundwater concentration is not an important issue if we are only interested in the short-term response to the change in recharge. However, over longer time periods this approach may not be appropriate. That is, the short-term response will be determined by the ambient groundwater concentration (i.e., the concentration of groundwater that was in the aquifer prior to the recharge event), but the long-term response will be determined by both the salinity of the recharge water and the salinity of the regional groundwater flow.

5.2.3 Transport Response

Different equations can be developed to estimate the movement of water directly (the *transport* process), and this approach is required if we wish to explicitly consider the salinity of the recharge water. Consider first the case where regional groundwater flow is very small, so that the flow to the river is generated principally by the recharge event. If the recharge event continues over time, then in the case of a long thin wetland parallel to the river (Case C) the time for the infiltrating water to move to the river is given by:

$$t_s = \frac{b\theta^2 x^2}{SH^2 K} \tag{5-5}$$

where t_s is the travel time for salt, θ is the aquifer porosity and H is the increase in watertable due to the recharge pulse (Welch *et al.*, 2013).[1] However, if recharge varies over time, then the travel time will be affected by the magnitude and distribution of recharge pulses in time, and so it is not possible to obtain such a simple expression for the travel time.

Alternatively, if the recharge pulse is very small, so that the regional groundwater gradient is not substantially changed (Cases A and B), then the travel time can be calculated as:

$$t_s = \frac{x\theta}{Ki} \tag{5-6}$$

where i is the regional hydraulic gradient.

If we consider steady flow of water between the point of recharge and the river, then the change in mass flux to the river arising from the recharge event is

$$\Delta m_r(t) = c(t - t_s)q(t - t_s) \tag{5-7}$$

where q is the recharge rate c is the concentration of wetland leakage. This equation simply says that the salt flux to the river is the concentration in recharge water at the time that infiltration occurred, multiplied by the infiltration rate at that time. Over time, the mass balance implied by Equation 5-7 will be correct (the total salt mass to the river will be equal to the total salt mass in recharge).

5.2.4 Mass Balance Correction

As noted, with appropriate choice of parameters, the mass balance is correct both for Equations 5-3 to 5-4 and 5-7. In other words, both the pressure equation and the transport equation individually account for all of the salt load to the river resulting from the increase in recharge. So, while it might appear superficially appealing to use Equation 5-3 to simulate the salt flux generated by the pressure response, and Equation 5-7 to simulate the

1 This equation results from a 1D solution for transport in a semi-infinite aquifer, following a sudden increase in head at the boundary (x=0). Derivation of the solution is given in Welch *et al.* (2013).

transport of salt, this will lead to a mass balance error, and the salt flux to the river will be double the salt flux in recharge. There is no simple solution to this problem, as these two approaches were not intended to be used together. Rather, they simulate part of the process, and are best used when only that part of the process is of interest.

Here, we trial an approach to represent both the pressure and transport processes within a single framework, while avoiding the mass balance error. To do this, Equation 5-3 is used to simulate the salt flux generated by the pressure response. We then use a modified form of Equation 5-7, designed to solve the mass balance problem.

$$\Delta m_r(t) = \big(c(t - t_s) - c_g\big)q(t - t_s) \tag{5-8}$$

The pressure equation is therefore used to calculate salt loads to the river, and the transport equation simply modifies this to simulate how salt concentrations change over time. Equation 5-8 is positive if the concentration in wetland leakage is greater than the ambient groundwater salinity and negative if the concentration in wetland leakage is less than the ambient groundwater salinity.

5.2.5 Regional Groundwater Flow

The above equations represent the change in water and salt flux to the river due to infiltration processes occurring on inundated areas of the floodplain. These equations modify the water and salt loads in regional groundwater flow. These are given simply as:

$$q_r(t) = q_g \tag{5-9}$$

and

$$m_r(t) = c_g q_g \tag{5-10}$$

where q_g is the regional groundwater flow rate towards the river and c_g is its concentration.

5.3 Implementation

5.3.1 Outline

This section develops simple numerical approaches for simulating the flow of water and salt through the groundwater system. Rather than develop a two-dimensional model of the floodplain, we focus on simple equations that use bulk parameters, such as areas of inundation, distance of inundated areas from the river, and average transport parameters. The aim is to apply these approaches for predicting changes in salt loads to the River Murray from infiltration beneath inundated areas of the river floodplain. Areas of inundation are represented as discrete wetlands. Although each wetland is represented using simple, average parameter values (e.g., wetland bed thickness and hydraulic conductivity), in principle, more complex patterns of inundation could be represented using multiple wetlands.

Equations for calculating the salt flux to the river are applied via the Source model (Welsh *et al.*, 2012). As part of its contribution to the eWater CRC, CSIRO developed functionality in Source that allows modelling of the water fluxes between groundwater and surface water in wetlands and floodplains (Jolly *et al.*, 2010; Rassam, 2011). A logical extension to this work is to develop capability within Source to predict the salinity accessions from wetlands/floodplains to rivers following overbank flows and environmental watering activities. This will allow salt generation processes by both groundwater and surface water to be modelled. Here, Equations 5-1 and 5-3 are used to simulate the water and salt fluxes, respectively, generated by the pressure response caused by wetland infiltration. The approach used for the transport response is analogous to the use of Equations 5-6 and 5-8, although a modification is made to also consider dispersion of solutes within the groundwater system. The use of Equation 5-6 implicitly assumes that the regional groundwater flow rate is large relative to the leakage rate beneath the wetland, and so the regional hydraulic gradient determines the travel time.

5.3.2 Formulation

The basic conceptualisation is illustrated in Figure 5-4. A wetland occurs adjacent to (but separated from) the river. A single channel joins the river and wetland, and is defined by a sill height. Surface water flow from the river to the wetland occurs when the river stage exceeds the sill level and the wetland water level. Return flow from the wetland to the river occurs when the wetland level exceeds both the sill level and the river level. Evaporation and infiltration remove water from the wetland and concentrate salt. The distance from the river bank to the centre of the wetland (*x*) and the area of the wetland (*A*) are key model parameters.

This simple conceptualisation is currently included in the Source model. However, water that infiltrates beneath the base of the wetland is currently lost to the model. In reality, though, this water (and dissolved salt) may move towards the river, to ultimately discharge into the river, with consequences for river salinity. We have modified a version of the Source model to include a groundwater flowpath from the wetland to the river.

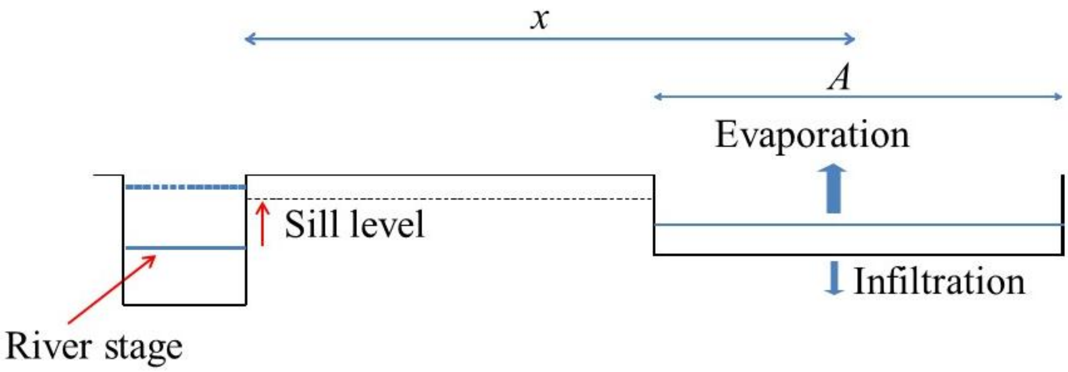


Figure 5-4 Conceptual model of river and connected wetland. Surface water flow from the river to the wetland occurs when the river stage exceeds the sill level and the wetland water level. Return flow from the wetland to the river occurs when the wetland level exceeds both the sill level and the river level. Evaporation and infiltration remove water from the wetland and concentrate salt. The distance from the river bank to the centre of the wetland (*x*) and the area of the wetland (*A*) are key model parameters.

In the current Source model, leakage through the wetland bed is given by:

$$q(t) = hC \tag{5-11}$$

where $q(t)$ is the infiltrative flow rate (m^3/day), C is the wetland bed conductance (m^2/day), and h is the water level in the wetland (m).

The daily time series for $q(t)$ becomes a recharge time series (after a time lag in the unsaturated zone, which is neglected in this formulation). The length of the recharge time series depends on the initial wetland volume and the rates of evaporation and infiltration, that is, the time series ends when h becomes zero (the wetland empties). From Equation 5-1, the discharge resulting from an instantaneous recharge source Q at time t_0 is given by:

$$q_r(t) = Q(t_0)\frac{x}{2(t-t_0)\sqrt{\pi D(t-t_0)}}\exp\left(\frac{-x^2}{4D(t-t_0)}\right) \tag{5-12}$$

where $q_r(t)$ is the instantaneous river discharge resulting from aquifer recharge sourced from wetland leakage (m^3/day), and $Q(t_0)$ is the instantaneous recharge sourced from the wetland bed during the daily time step (m^3). A convolution approach is adopted to sum the impacts of the daily recharge pulses to yield cumulative discharge to the river at any time. The salt flux associated with this discharge is simply given by

$$m_r(t) = c_g\, q_r(t) \tag{5-13}$$

where c_g is the groundwater salinity adjacent to the river.

Similarly, the salt flux infiltrating through the base of the wetland that discharges to river after a time lag, can be expressed:

$$m_r(t) = \left(c \quad (t_0) - c_g\right)Q(t_0)\frac{x}{2(t-t_0)\sqrt{\pi D_s(t-t_0)}}\exp\left\{\frac{-[x-v(t-t_0)]^2}{4D_s(t-t_0)}\right\} \qquad (5\text{-}14)$$

where $m_r(t)$ is the daily salt mass arriving at the river resulting from the daily salt pulse that has infiltrated the wetland bed, *v* is the groundwater velocity, and D_s is the solute dispersivity (Maloszewski and Zuber, 1982; Zuber, 1986). This equation is essentially the same as Equation 5-7, except that dispersion of solute occurs during flow to the river. [2]

Equations 5-12 and 5-13 represent the pressure response from the infiltration, and Equation 5-14 represents the transport of salt from wetland to the river. The use of both equations allows us to both simulate the pressure response, and to explicitly consider how the concentration of salt in groundwater flowing into the river changes over time. A convolution approach is adopted to sum the impacts of the daily salt load pulses to yield cumulative salt discharge to the river at any time.

Key parameters required by the model are:

- Regional groundwater velocity. (This is determined in the model from the regional hydraulic gradient, hydraulic conductivity and porosity.)
- Wetland area
- Evaporation rate from wetland
- Wetland bed thickness and conductance
- Solute dispersivity

Table 5-1 List of symbols used

Symbol	Description	Units	Symbol	Description	Units
A	Area of wetland	L^2			
b	Aquifer thickness	L	L	Leakage rate	L/T
c	Concentration in wetland	M/L^3	M	Salt mass in the wetland	M
c_g	Ambient groundwater concentration	M/L^3	m_r	Salt flux to the river	M/T
c_{riv}	Concentration in river	M/L^3	M_r	Salt flux to the river	M
C	Floodplain bed conductance	L^2/T^1	q	Recharge/leakage rate to aquifer	L^3/T
D	Hydraulic diffusivity	L^2/T^1	q_r	Groundwater discharge to the river	L^3/T
Ds	Solute dispersivity	L	Q	Leakage flux beneath wetland	L^3
E	Floodplain evaporation rate	L/T	Q_r	Cumulative discharge to the river	L^3
h	Water depth in wetland	L	S	Specific yield	-
h_c	Thickness of wetland bed	L	V	Groundwater velocity	L/T
H	Change in groundwater level	L	x	Distance of wetland from the river	L

[2] This is essentially the Dispersive Model, described by Maloszewski & Zuber (1982). Further derivation is provided in this paper.

Symbol	Description	Units	Symbol	Description	Units
i	Hydraulic gradient	-			
K	Aquifer hydraulic conductivity	L/T	θ	Aquifer porosity	-
K_c	Hydraulic conductivity of wetland bed	L/T			

5.3.3 Example 1: Separate pressure and transport peaks

5.3.3.1 Outline

A simple model was developed in order to test the efficacy of the groundwater flow and solute transport processes added to the Source software. In this example, model parameter and boundary condition values were manipulated in order to achieve results in which separate pressure and transport responses were readily identifiable. In practice, these two components of a solute breakthrough curve will typically be combined within a single groundwater solute response and therefore not readily distinguishable from one another. Note that the model parameters used in these simulations are not intended to represent the River Murray or any other river - wetland system.

5.3.3.2 Conceptual model

The model simulates the movement of water and solute from a fully inundated wetland through an aquifer to a river (Figure 5-5). In Source terminology, this is represented using two nodes (i.e. one storage node and one gauge node) and one link (i.e. one groundwater routing-type link). A temporal extent of 11 years (i.e. 2000–2011) was sufficient for model testing purposes.

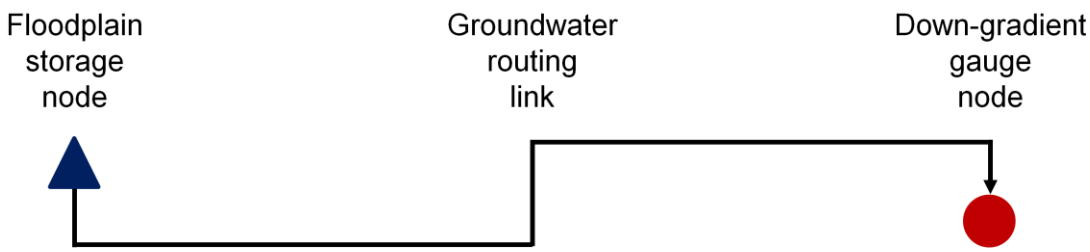


Figure 5-5 Conceptual model used to investigate separate pressure and transport responses simulated using the modified version of Source

Table 5-2 Model parameter values.

Model element	Parameter	Value	Units
Wetland storage node	bed conductance	100	m^2/day
	wetland area	10^6	m^2

Model element	Parameter	Value	Units
	initial water volume	100	ML
	initial solute concentration	2,000	mg/l
Groundwater storage link	aquifer hydraulic conductivity	10	m/day
	aquifer thickness	10	m
	aquifer porosity	10	%
	aquifer specific yield	0.1	-
	hydraulic gradient	5×10^{-4}	-
	orthogonal distance to discharge location	100	m
	solute dispersivity	0.1	m
	solute concentration	1,000	mg/l

The wetland (i.e. storage) node served as the up-gradient boundary condition and featured a maximum (and initial) volume of 100 ML distributed over an area of 1 km^2. This corresponded to a maximum stage height of 10 m. The initial concentration of wetland water was specified as 2,000 mg/l and the bed conductance was specified as 100 m^2/day. The groundwater routing link was parameterised as shown in Table 2, including the initial ambient groundwater concentration, which was specified as 1,000 mg/l.

5.3.3.1 Results

Modelling results are presented in terms of water mass fluxes, solute mass fluxes, and solute concentrations both exiting the wetland and entering the river (Figure 5-6). The rate of water infiltration though the base of the wetland to groundwater commences at 1 ML/day and declines exponentially to zero after approximately one year. By the end of the first day, the rate of groundwater discharge to the river is approximately 0.1 ML/day. As the propagation of hydraulic pressure resulting from wetland discharge reaches the far extent of the aquifer, the rate of groundwater discharge rises to ~0.5 ML/day after approximately one month. The rate of groundwater discharge declines thereafter.

The rate of solute mass infiltrated though the base of the wetland to groundwater commences at 2 tonnes/day and declines exponentially to zero after ~1.5 years elapsed. The rate of groundwater solute mass flux to the river after one day is approximately 0.07 tonnes/day. The solute mass transported by the hydraulic pressure response can be observed in the peak rate of ~0.6 tonnes/day occurring after approximately one month. This response persists for approximately 1.5 years. Conversely, increases in solute mass flux due to dispersive transport (i.e. the transport response) may be observed from the year ~2003 onward. The rate of solute mass flux increases to ~0.1 tonnes/day after ~5.5 years elapsed before decreasing to zero after ~9 years elapsed.

The solute concentration of water exiting the wetland (primary axis) remains constant at 2 g/l until wetland infiltration ceases at ~4.5 years elapsed. The solute concentration of groundwater discharge (secondary axis) is calculated by dividing the solute flux by the water flux, and increases from 1 g/l (the concentration of ambient groundwater) to a maximum of ~100 g/l after ~6 years before reducing towards 1 g/l by 2011. It is important to note that the transport solute mass peak is not associated with a water mass peak. The transport equation simulates a solute flux but does not explicitly simulate a water flux. For this reason, concentrations associated with the transport peak are not necessarily meaningful. Nevertheless, the concentration anomalies are unlikely to be observed when river concentrations are considered.

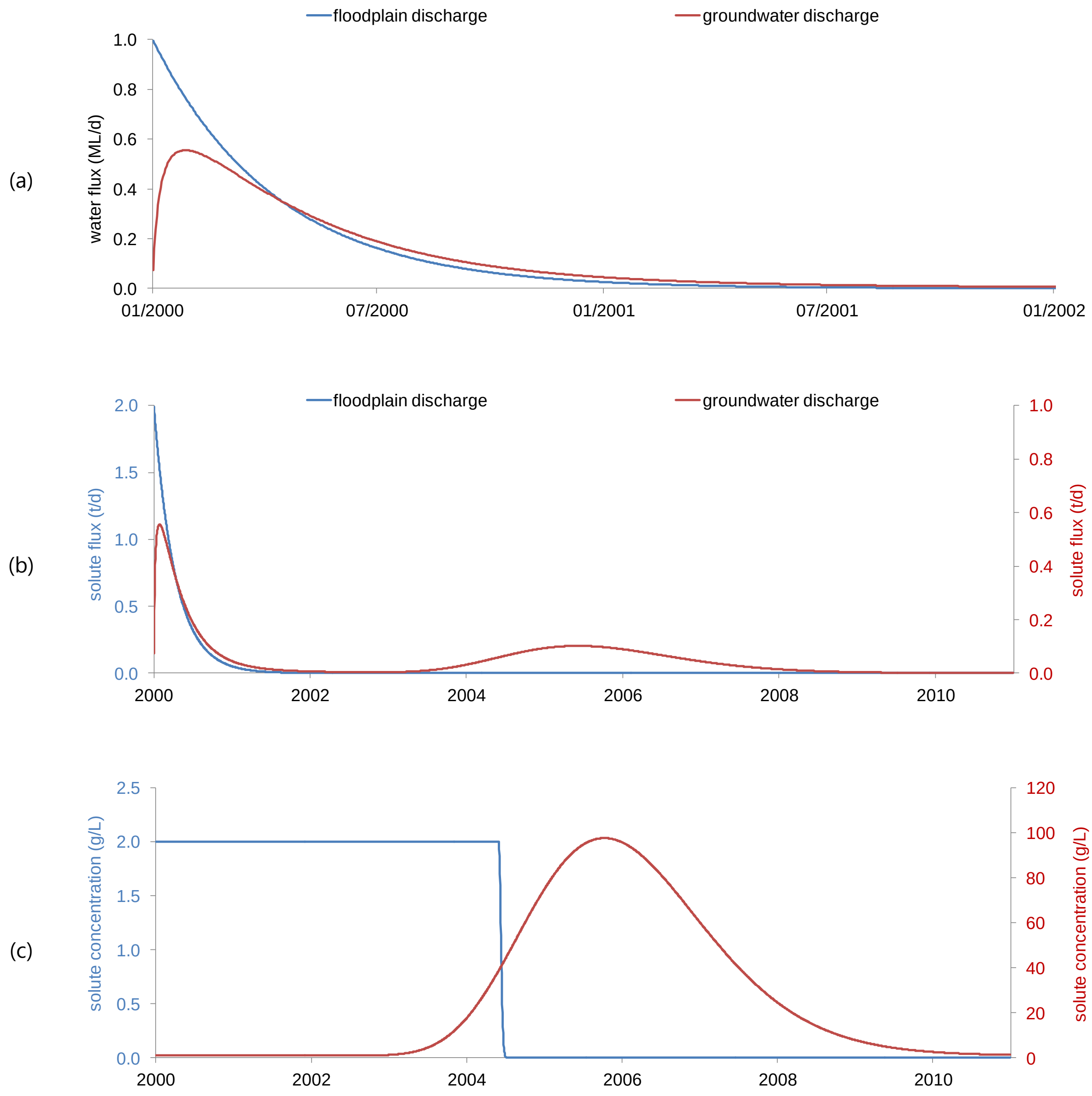


Figure 5-6 (a) Water mass fluxes, (b) solute mass fluxes and (c) solute concentrations of water exiting the wetland (blue) and groundwater (red). Note that the time scale is expanded for (a).

5.3.3.2 Sensitivity analyses

A simple sensitivity analysis was undertaken to examine how water flux and salt load varied as model parameters were varied. In the following, each relevant model parameter was increased and decreased by a factor of two.

The sensitivities of water mass fluxes exiting groundwater to parameter variation are presented in terms of water mass fluxes, solute mass fluxes, and solute concentrations (Figures 5–10).

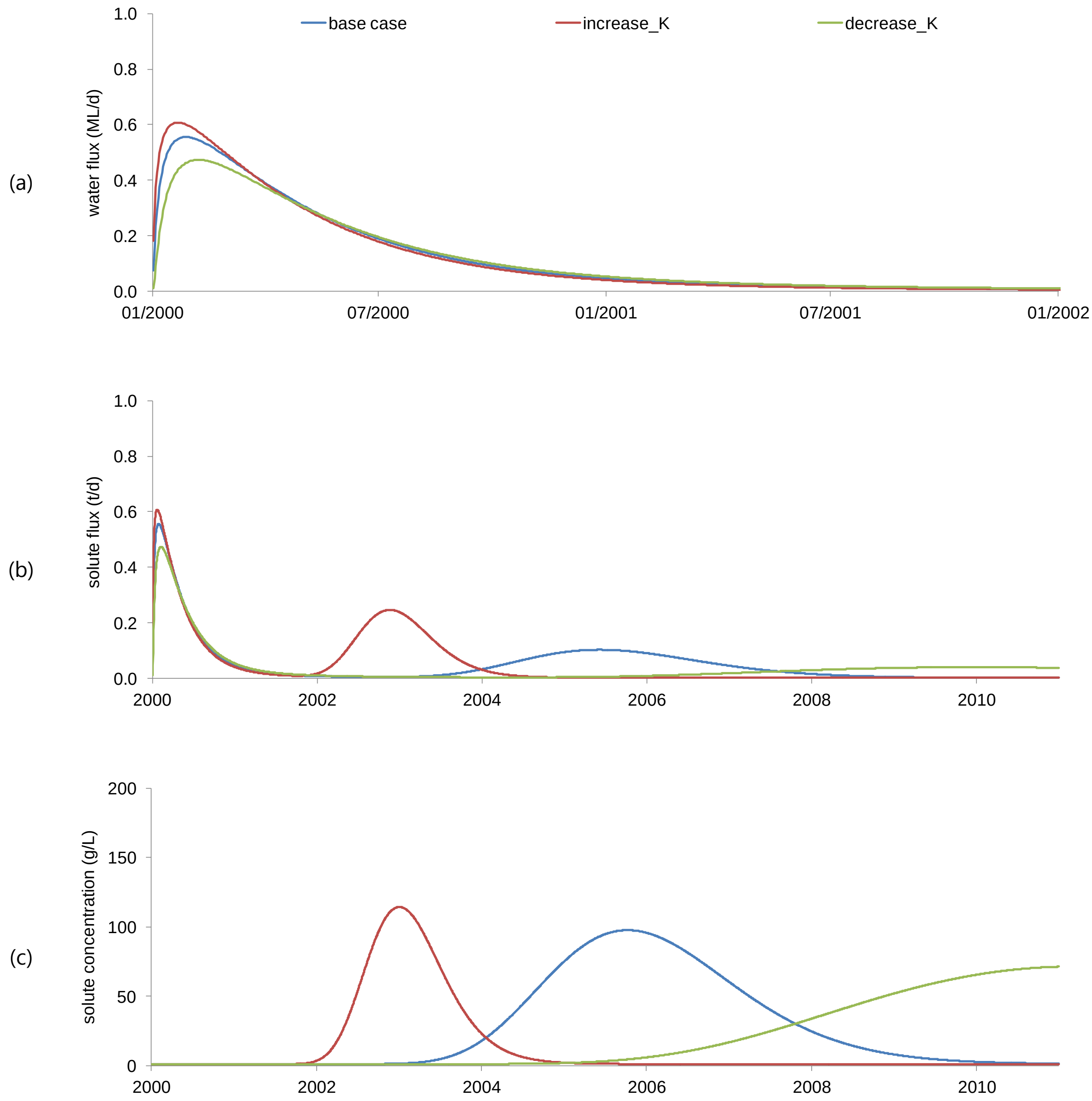


Figure 5-7 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of groundwater discharge to changes in aquifer hydraulic conductivity. Note that the time scale is expanded for (a).

Both the pressure and transport responses are sensitive to aquifer hydraulic conductivity (Figure 5-7 and Equations 5-12 and (5-15). (Although K does not explicitly appear in either equation, it is involved in calculation of the D and v terms in these equations.) Thus changes in aquifer hydraulic conductivity affect the peak water flux and its arrival time (Figure 5-7a). Both solute mass peaks are affected by hydraulic conductivity, with shorter response times and larger salt mass peaks produced by greater aquifer hydraulic conductivity values (Figure 5-7b and c).

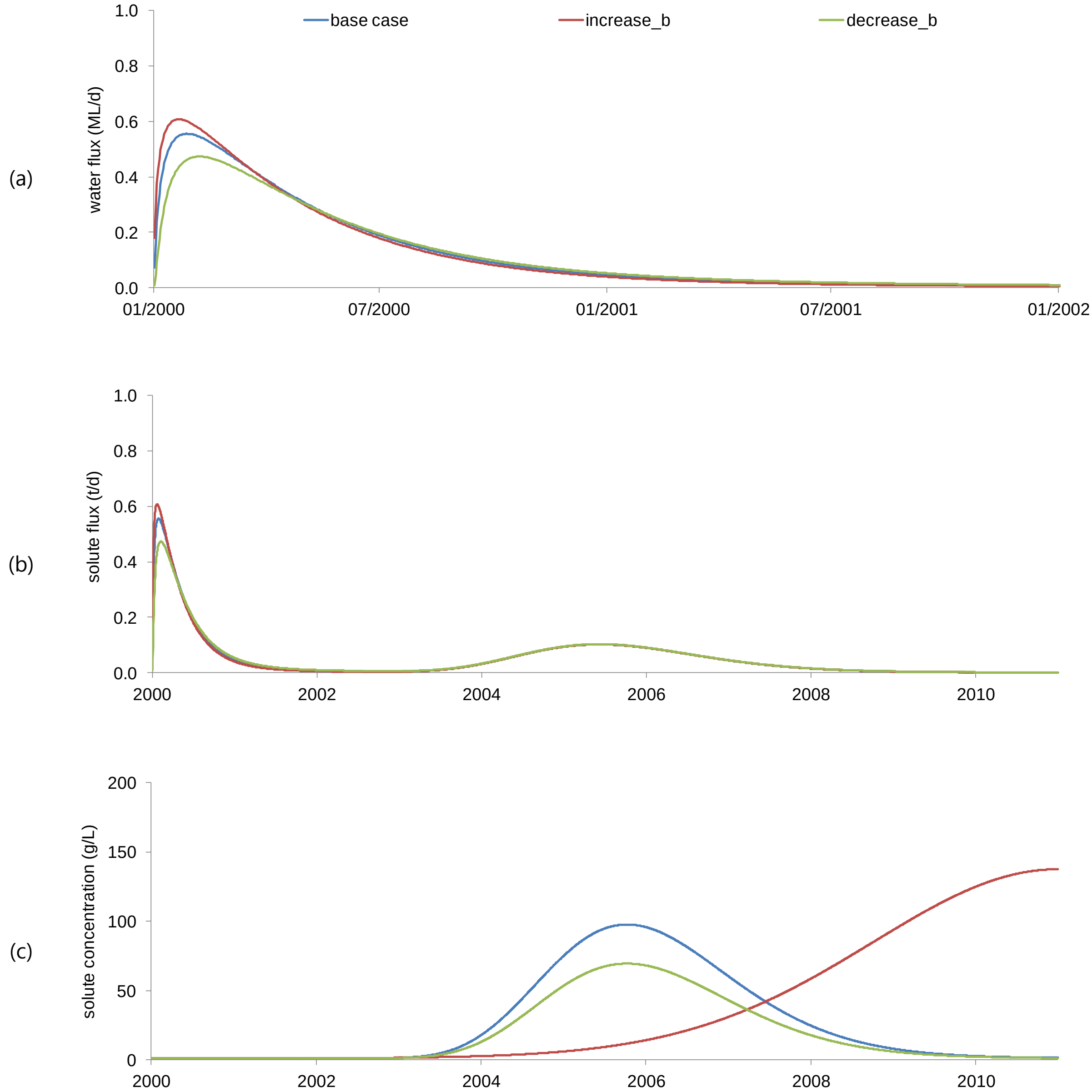


Figure 5-8 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of groundwater discharge to changes in aquifer thickness. Note that the time scale is expanded for (a).

Changes in aquifer thickness affected calculation of aquifer diffusivity (pressure response; Equation 5-12), and so affect peak water flux (Figure 5-8a) and the arrival time and magnitude of the first solute mass flux (Figure 5-8b - note the difference at very early times). The second solute mass peak is independent of the aquifer thickness. As discussed above, solute concentration is simply the ratio of salt mass flux to water flux. Higher concentrations (Figure 5-8c) are therefore due to lower water mass fluxes (Figure 5-8a).

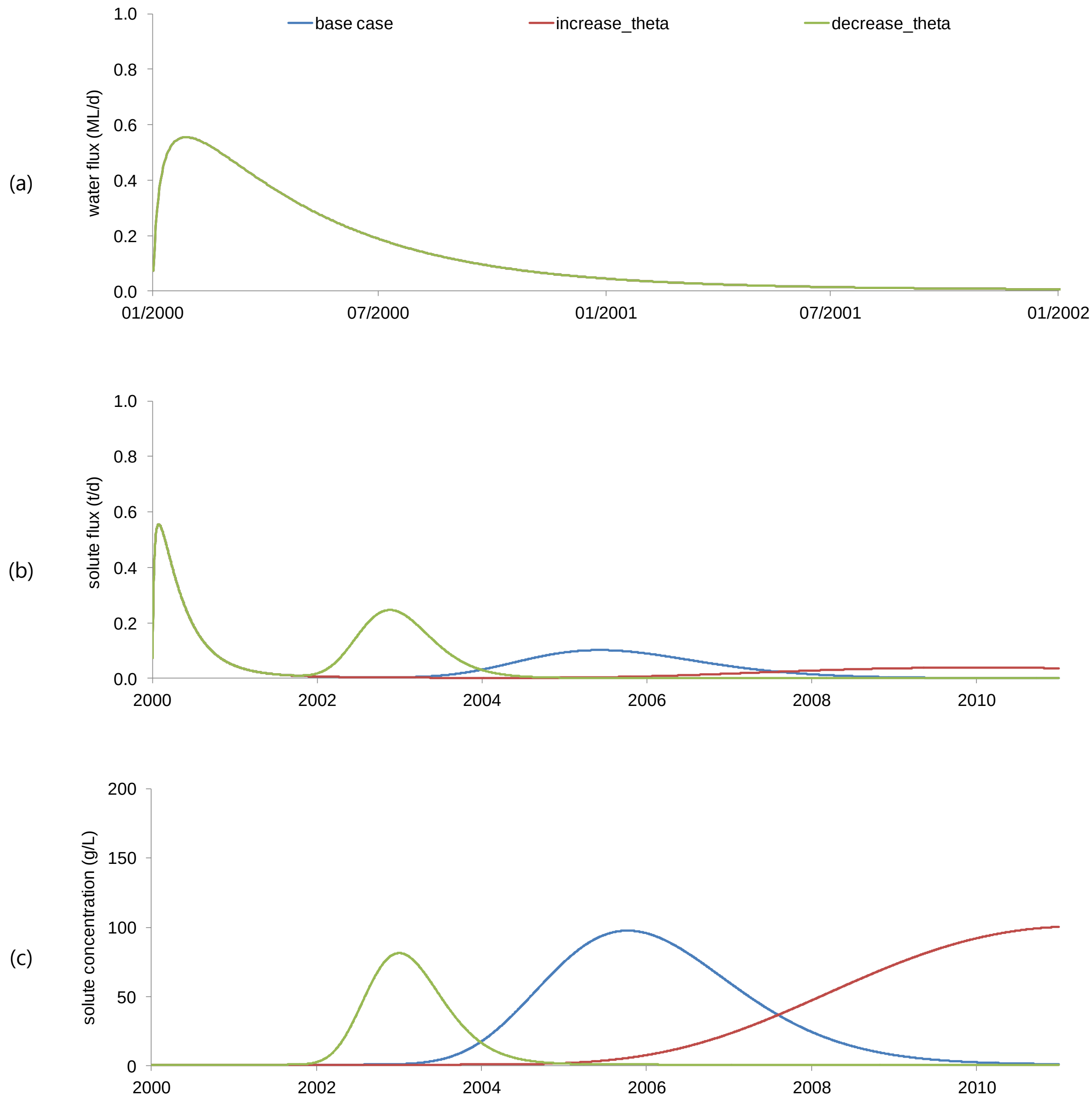


Figure 5-9 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of discharge to changes in aquifer porosity. Note that the time scale is expanded for (a).

Changes in aquifer porosity do not affect the pressure response (Figure 5-9a), as this term does not appear in Equation 5-12. Aquifer porosity, however, affects the groundwater velocity (Equation 5-15). Increases in porosity cause a decrease in velocity, and hence longer transport response times (Figure 5-9b).

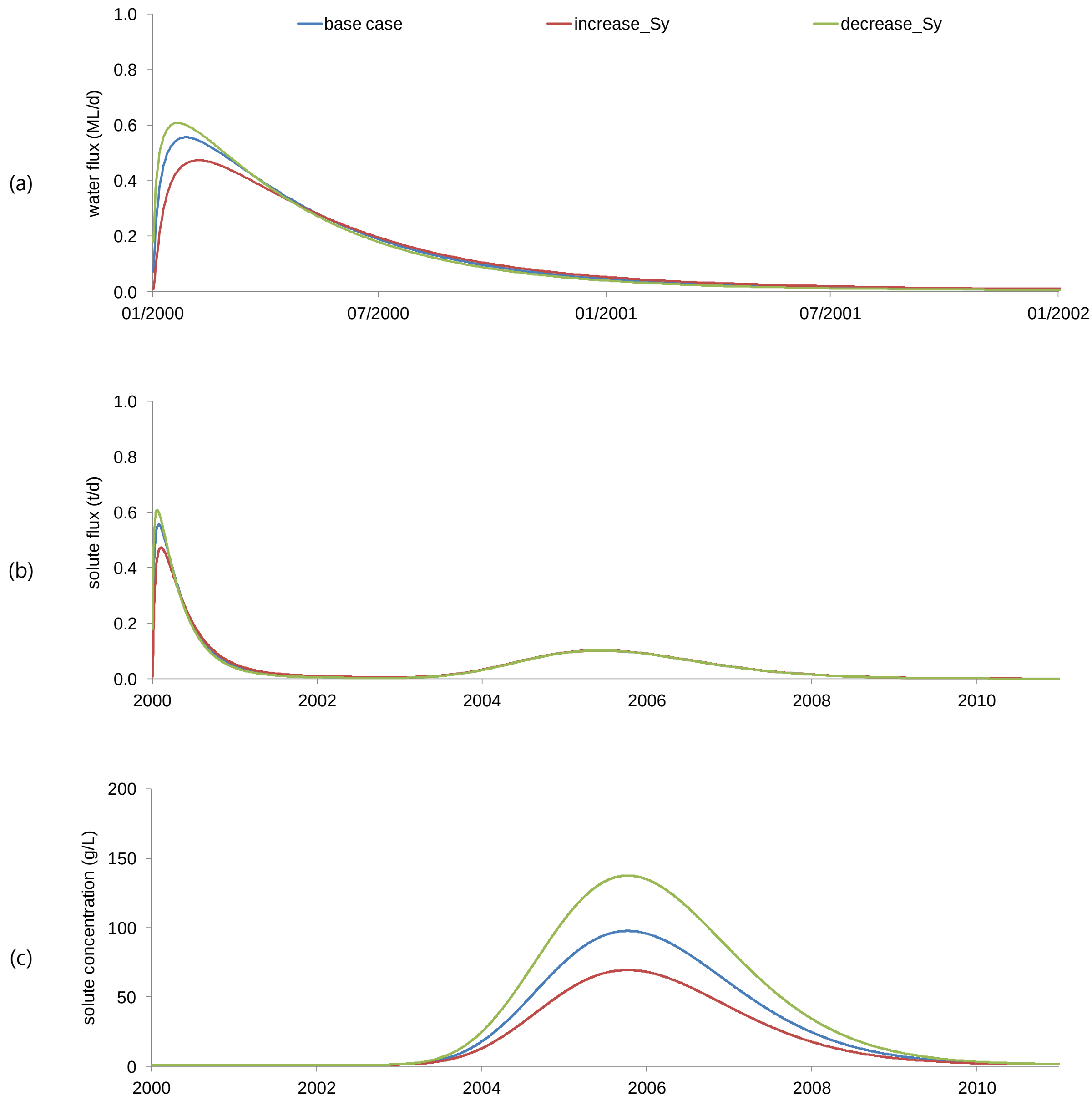
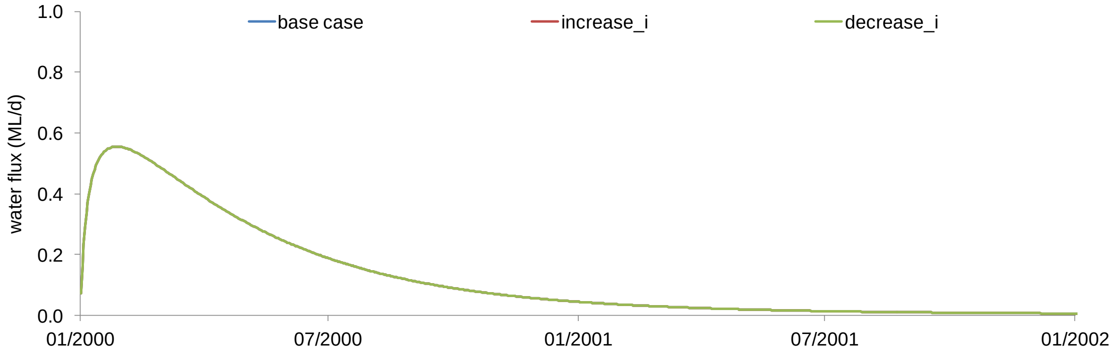


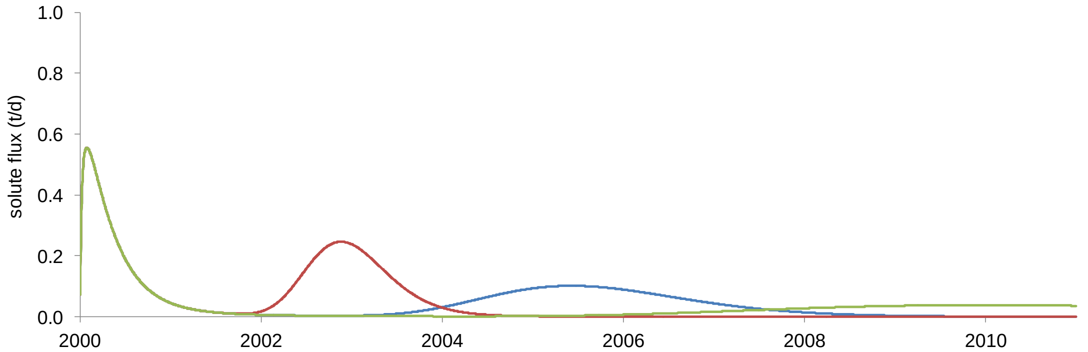
Figure 5-10 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of groundwater discharge to changes in aquifer specific yield. Note that the time scale is expanded for (a).

Changes in aquifer specific yield affect the pressure response but not the transport response. Aquifer diffusivity is inversely proportional to specific yield, and so doubling the specific yield produces the same effect as halving aquifer thickness (compare Figure 5-10 and Figure 5-8).

(a)



(b)



(c)

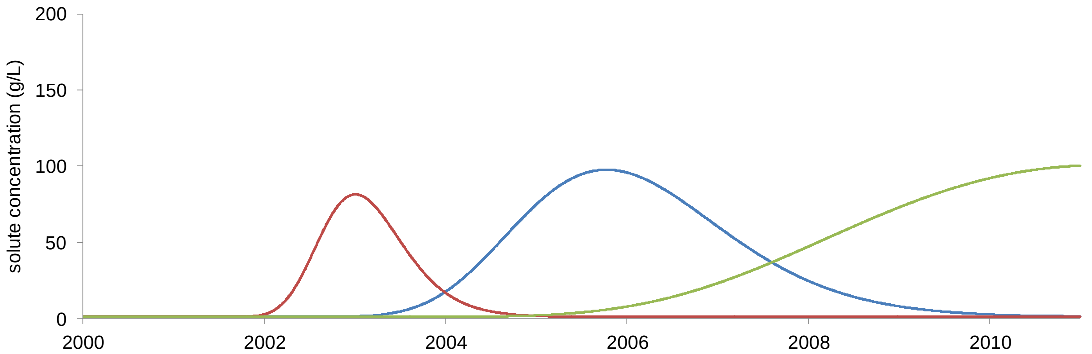


Figure 5-11 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of groundwater discharge to changes in groundwater hydraulic gradient. Note that the time scale is expanded for (a).

Changes in hydraulic gradient affect the transport response, but not the pressure response (Figure 5-11a and b). An increase in hydraulic gradient reduces the transport travel time (Figure 5-11b). Doubling hydraulic gradient has the same effect as halving the porosity (compare Figure 5-11 and Figure 5-9).

Changes in aquifer solute dispersivity affect the transport response, but not the pressure response (Figure 5-12a and b). Increasing solute dispersivity reduces the salt peak associated with the transport process and reduces the transport travel time (Figure 5-12b).

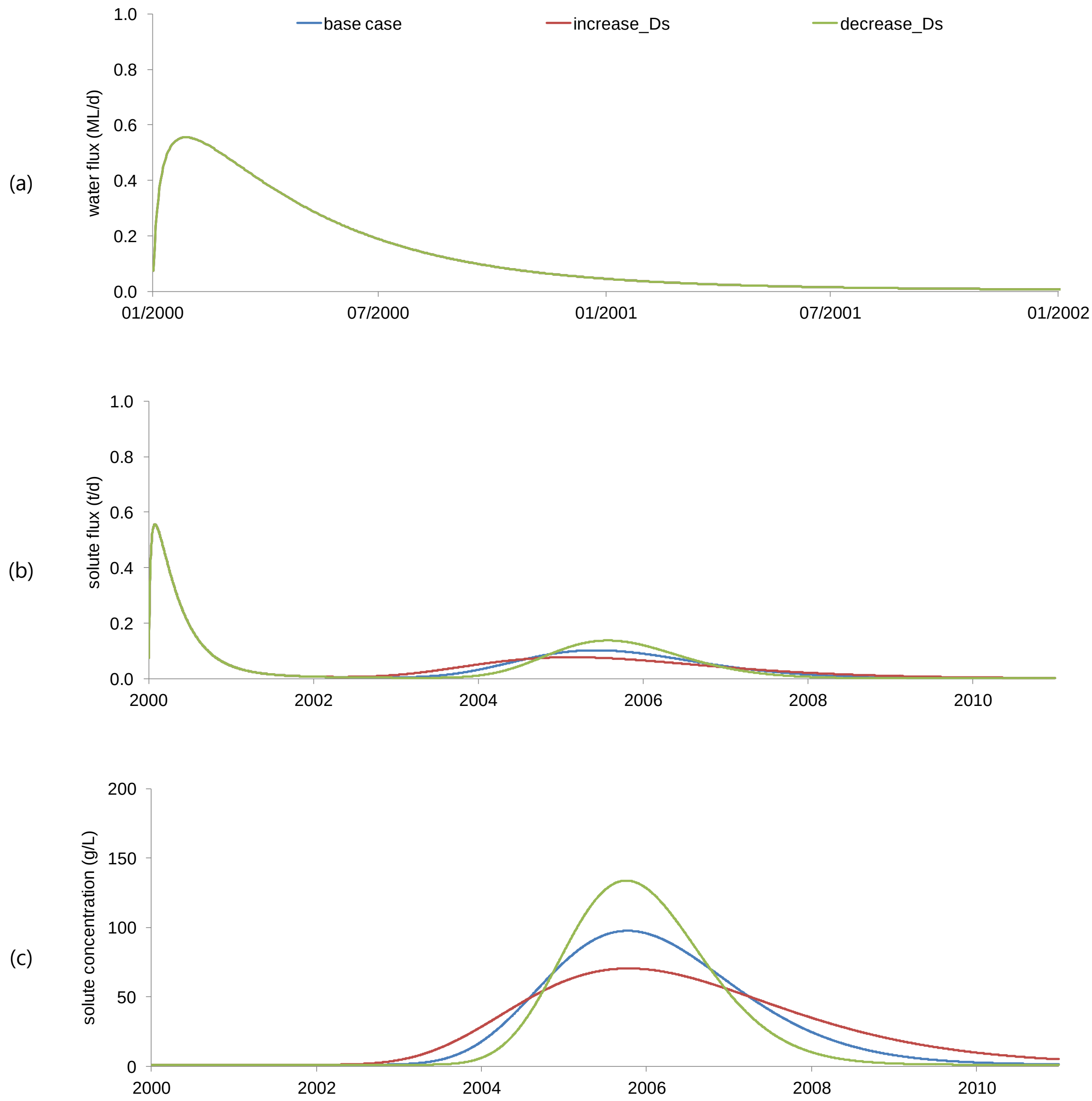


Figure 5-12 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of groundwater discharge to changes in aquifer solute dispersivity. Note that the time scale is expanded for (a).

5.3.3.3 Potential for negative mass fluxes

The calculation of solute transport by dispersion (Equation 5-15) involves the subtraction of the ambient groundwater concentration. This raises the possibility for the calculation of negative concentrations and negative solute fluxes for water discharged to the river. The potential for this problem was examined by assigning an extremely high specified groundwater concentration (100,000 mg/l) and re-running the model. The solute mass discharged from groundwater is presented in Figure 5-13.

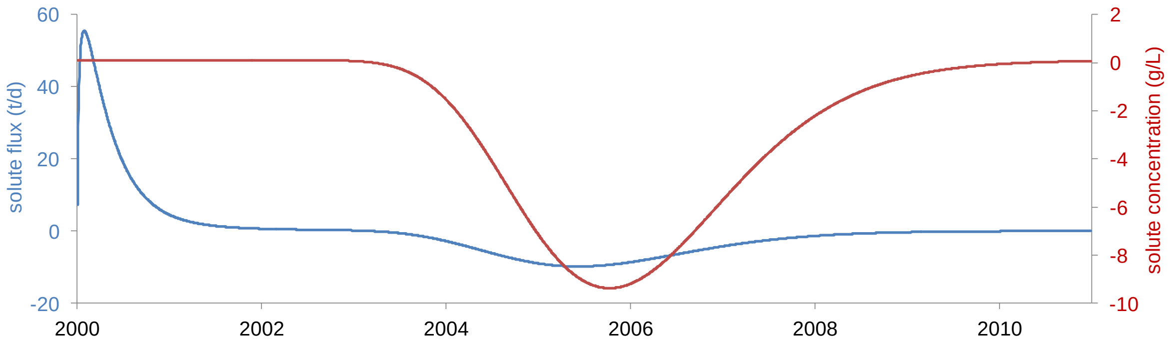


Figure 5-13 Solute mass (blue) and solute concentration (red) discharged from groundwater for scenario featuring high specified groundwater concentration.

The high specified groundwater concentration results in the calculation of negative solute mass fluxes. (In fact, negative values will occur whenever $c_g > c_w$.) This may be observed in the reductions in groundwater solute mass flux and concentration from 2003 to ~2006. These metrics do not return to physically realistic (i.e. nonzero) values until approximately 5 years have elapsed. Although calculations using Equation (?) do ensure that solute mass balance will always be preserved, as shown, they do present potential for errors in the calculation of solute mass fluxes and concentrations in groundwater discharged to a river.

5.3.4 Example 2: Draining wetland model

This example was designed to test the performance of the modified Source software code (a) when realistic parameter values were specified and (b) when regional groundwater flow was represented.

5.3.4.1 Conceptual model

Compared to the model used previously to identify separate solute breakthrough concentration curves for groundwater pressure and transport responses, the model used for this example includes some key changes. First, an additional node and link were added, which enabled the inclusion of an additional specified groundwater flux at a specified solute concentration. This additional flux represents regional groundwater flow. The groundwater link parameter values were also changed in order to be more realistic and to be consistent with values typically used when simulating groundwater–wetland interactions on the lower River Murray. In Source terminology, this example uses three nodes (i.e. one storage node and two gauge nodes) and two links (i.e. one groundwater routing-type link ["Groundwater link 1"] and one storage routing-type link ["Groundwater link 2"]) (Figure 5-14). A temporal extent of 51 years (i.e. 2000–2050) was sufficient for model testing purposes.

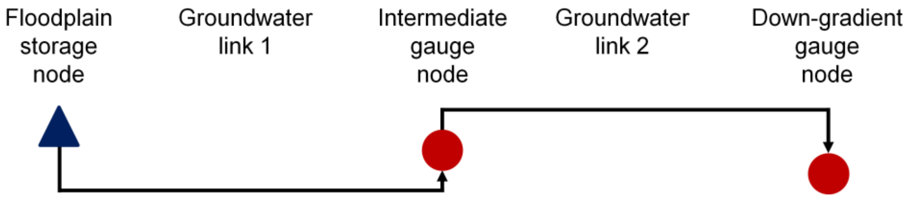


Figure 5-14 Conceptual model used to test the performance of the modified Source software code (a) when realistic parameter values were specified and (b) when regional groundwater flow was represented.

5.3.4.2 Parameterisation

The wetland (i.e. storage) node served as the up-gradient boundary condition and featured an initial volume equal to its maximum (and initial) volume of 5,000 ML, distributed over an area of 1 km^2. This corresponded to a maximum stage height of 5 m. The initial concentration of wetland water was specified as 10,000 mg/l and the bed conductance was specified as 20,000 m^2/day. The bed conductance value was manipulated in order to achieve a maximum linear water flux from the wetland of 0.1 m/day, which is comparable with those measured by Bramley *et al.* (2003). An evaporation flux from the wetland was also included, at a rate of 5 mm/day. This value is consistent with the mean annual evaporation observed at the Loxton Research Centre (http://www.bom.gov.au) as well as with values observed at Calperum Station (Meyer *et al.,* 2015).

The spatial extent of the wetland was selected arbitrarily, although it is worth noting that fluxes from this area of floodplain are added to the river at a point. Thus, the calculated fluxes are not associated with any particular length of river. Rather, different areas of inundation (in terms of square metres of wetland per kilometre of river) can be simulated either by changing the wetland area or by changing the frequency with which wetlands are attached to the river.

The groundwater routing link (hereafter "groundwater link 1") was parameterised as follows. The aquifer simulated featured hydraulic conductivity = 20 m/day, thickness = 20 m, porosity = 0.3, specific yield = 0.15. These parameter values are considered representative of the Monoman Formation surficial sand aquifer (Woods *et al.*, 2014, Woods, 2015a). Other aquifer properties, such as hydraulic gradient = 1 x 10^{-3}, orthogonal distance to discharge location = 100 m, and solute dispersivity = 1 m, were selected arbitrarily but are considered physically reasonable. In addition, groundwater was assigned a constant concentration of 10,000 mg/l, which is consistent with that observed within the floodplain (See Figure 3-25 in Woods (2015a).

For the storage routing-type link located immediately down-gradient of the previously discussed groundwater link (hereafter "groundwater link 2"), a constant groundwater discharge of 20 m^3/day was specified, featuring a constant concentration of 10,000 mg/l.

All parameter values specified for the model are summarised in Table 5-3.

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Table 5-3 Model parameter values

Model element	Parameter	Value	Units
Wetland storage node	bed conductance	20,000	m^2/day
	wetland area	10^6	m^2
	initial water volume	5,000	ML
	initial solute concentration	10,000	mg/l
Groundwater storage link	aquifer hydraulic conductivity	20	m/day
	aquifer thickness	20	m
	aquifer porosity	30	%
	aquifer specific yield	0.15	-
	groundwater hydraulic gradient	1×10^{-3}	-
	orthogonal distance to discharge location	100	m
	solute dispersivity	1	m
	solute concentration	10,000	mg/l

5.3.4.3 Results

Modelling results are presented in terms of water mass fluxes, solute mass fluxes, and solute concentrations (Figure 5-15). It should first be noted that, due to the parameter values specified in this test case, separate pressure and transport responses cannot be distinguished from one another; instead, both responses occur on similar timescales. The rate of water infiltration though the base of the wetland to groundwater commences at 100 ML/day and declines exponentially to zero after approximately one year elapsed. The rates of discharge from groundwater links 1 and 2 (which are almost identical) commence at ~10 ML/day. As the propagation of hydraulic pressure resulting from wetland discharge reaches the far extent of the aquifer, the rate of groundwater discharge rises to ~55 ML/day at ~3 months elapsed. The rate of groundwater discharge declines exponentially thereafter. Note that the time series of discharge rate from groundwater link 2 is almost identical to that of groundwater link 1. This is because the specified additional influx due to regional groundwater flow was insignificant (0.02 ML/day) in comparison to that resulting from wetland infiltration.

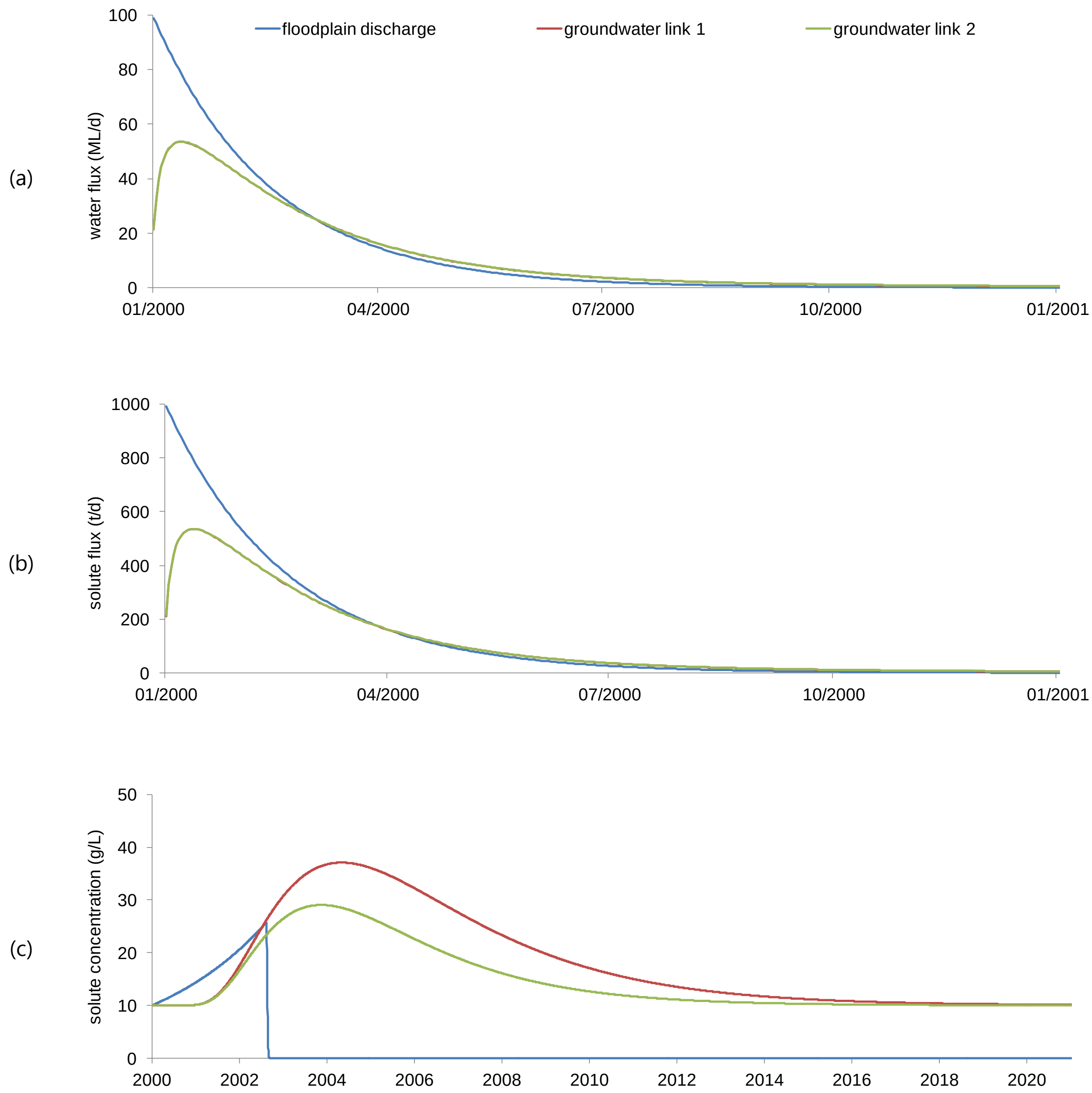


Figure 5-15 (a) Water mass fluxes, (b) solute mass fluxes and (c) solute concentrations of water exiting the wetland (blue), groundwater link (red) and groundwater link 2 (green). Note that the time scale is compressed for (c).

The rate of solute mass infiltrated though the base of the wetland to groundwater commences at 1,000 tonnes/day and declines exponentially to zero after ~1 years elapsed. The rates of solute mass flux from both groundwater links commence at ~200 tonnes/day. The solute mass transported can be observed in the peak rate of ~550 tonnes/day occurring after less than three months elapsed. This response, which combined both pressure and transport responses, persists until ~1 year has elapsed.

The solute concentration of water exiting the wetland increased from ~10,000 mg/l to ~25,000 mg/l at ~2.5 years elapsed, at which time infiltration from the wetland ceases. This increase in wetland solute concentration is due to evapotranspiration. The solute concentration of discharge from groundwater link 1 increases from 10,000 mg/l to a maximum of ~35,000 mg/l after ~4 years elapsed. Similarly, the solute concentration of discharge from groundwater link 2 increases from 10,000 mg/l to a maximum of ~25,000 mg/l after ~4 years elapsed. As discussed above, these high concentrations result from slight differences in the temporal patterns of the pressure and transport responses to infiltration. This means that water and salt fluxes can occur at different times, resulting in anomalous concentrations when salt flux is divided by water flux. However, these anomalies are unlikely to be observed when river concentrations (rather than groundwater discharge concentrations) are considered.

5.3.4.4 Sensitivity analyses

In order to test the sensitivity of the representations of groundwater flow and solute transport to parameter variation, the one-at-a-time approach was again undertaken (Figure 5-16 – Figure 5-21). For the parameter values examined, changes in aquifer hydraulic conductivity appear to be nonlinearly related to peak water and solute fluxes and the arrival times thereof (Figure 5-16a and b). For example, in comparison to the base case scenario, increases and decreases in hydraulic conductivity result in reductions in both peak water and solute mass fluxes. Peak solute concentration is positively correlated with aquifer hydraulic conductivity, while the timing of this response is inversely correlated (Figure 5-16c).

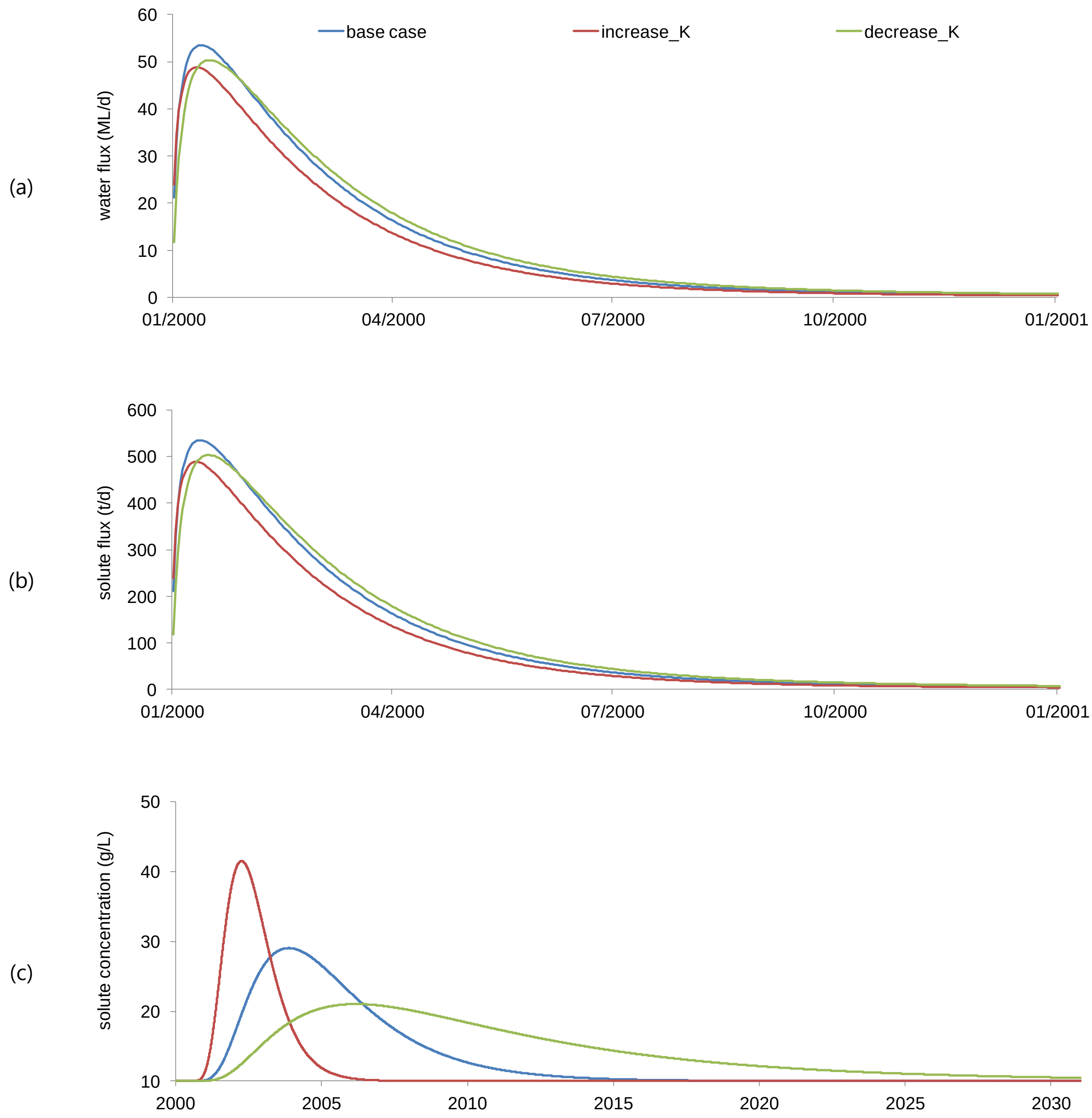


Figure 5-16 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of groundwater discharge to changes in aquifer hydraulic conductivity. Note that the time scale is compressed for (c).

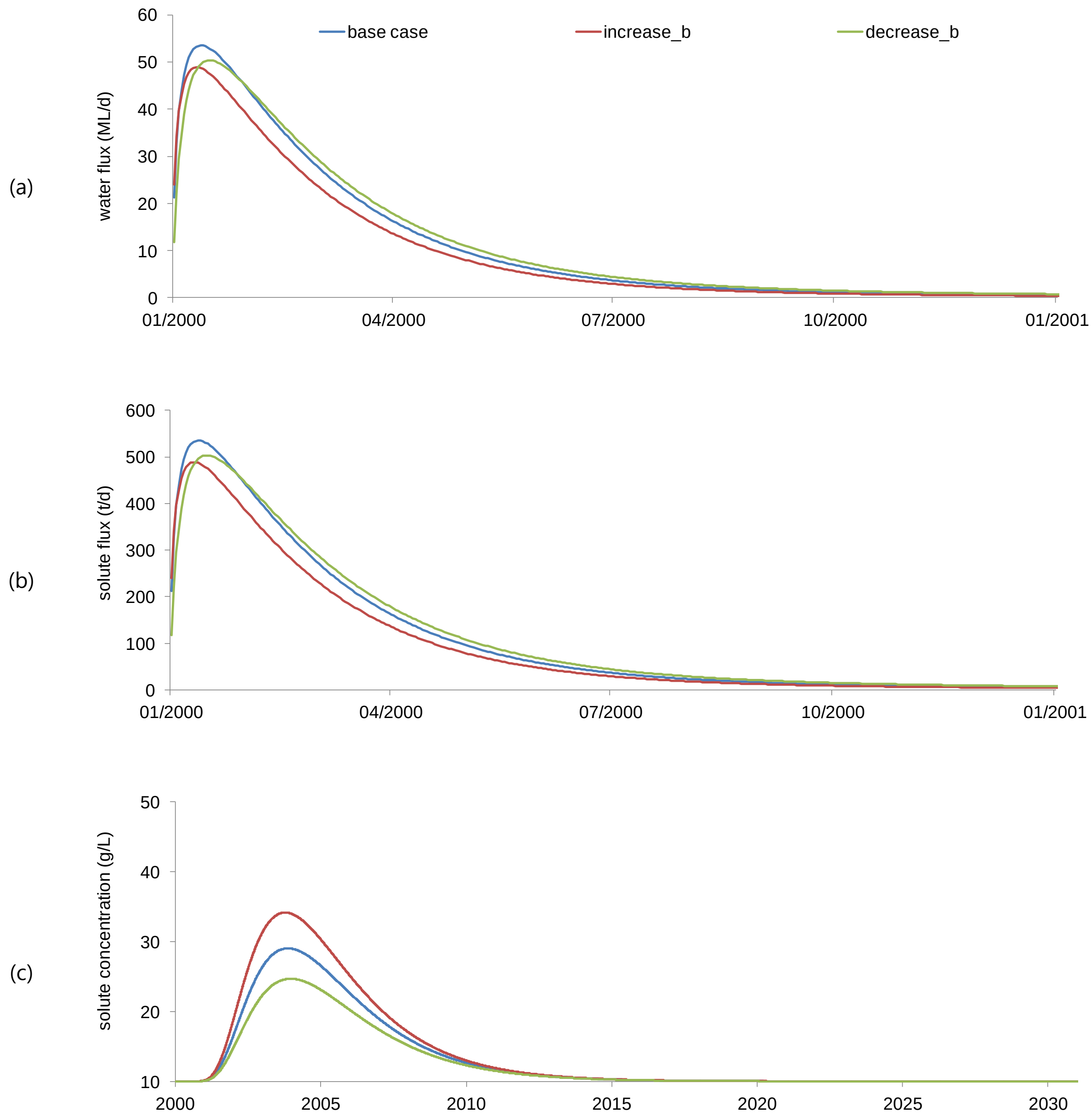


Figure 5-17 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of groundwater discharge to changes in aquifer thickness. Note that the time scale is compressed for (c).

For the parameter values examined, changes in aquifer thickness appear to be nonlinearly related to peak water and solute fluxes and the arrival times thereof (Figure 5-17a and b). For example, in comparison to the base case scenario, increases and decreases in aquifer thickness result in reductions in both peak water and solute fluxes. Peak solute concentration is positively correlated with aquifer thickness, while the timing of this response is insensitive to variations in aquifer thickness (Figure 5-17c).

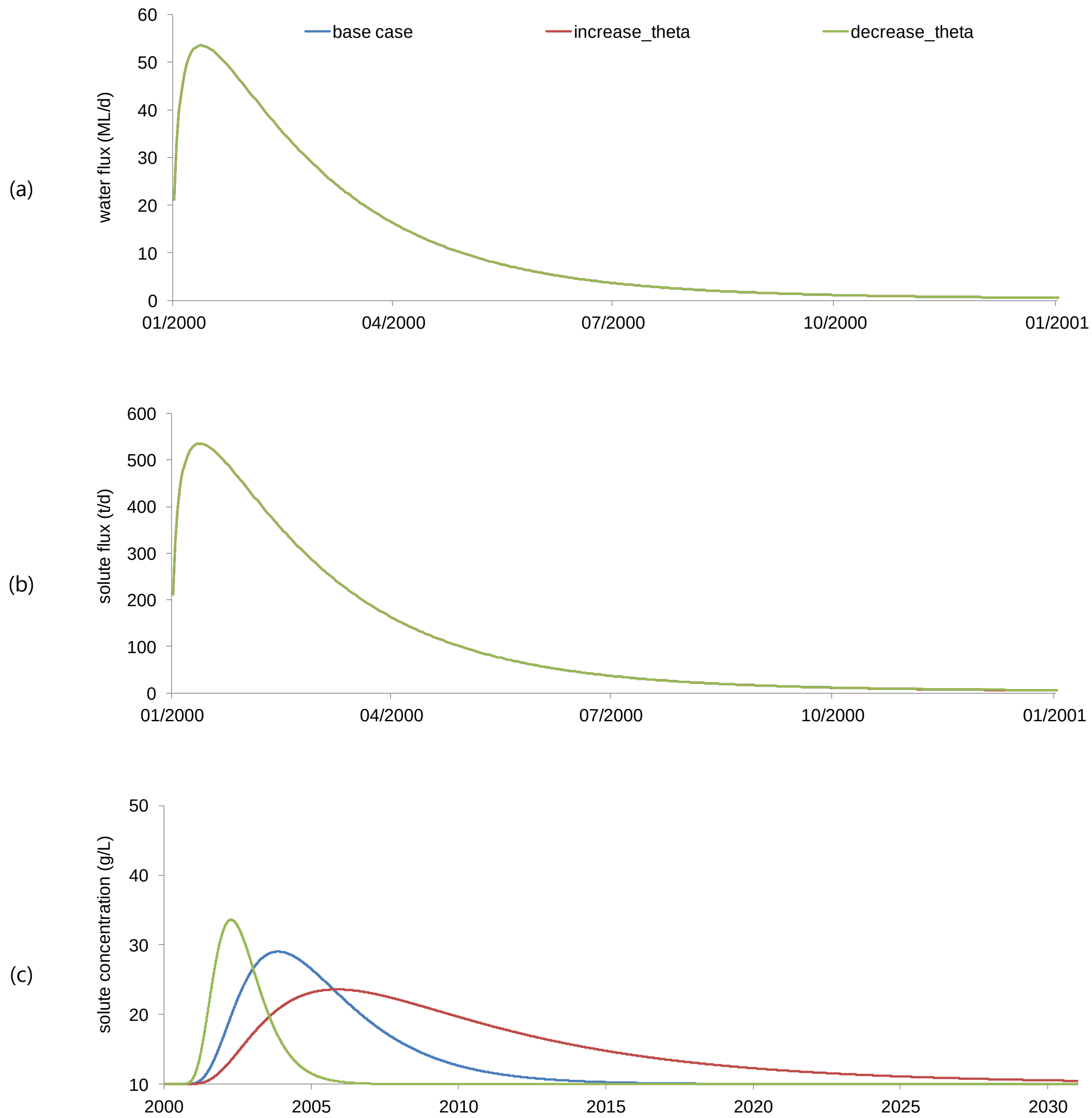


Figure 5-18 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of discharge to changes in aquifer porosity. Note that the time scale is compressed for (c).

Changes in aquifer porosity have minimal effect on both water and solute fluxes and the timing of these responses (Figure 5-18a and b). Peak solute concentration is inversely correlated with porosity, while the timing of this response is positively correlated (Figure 5-18c).

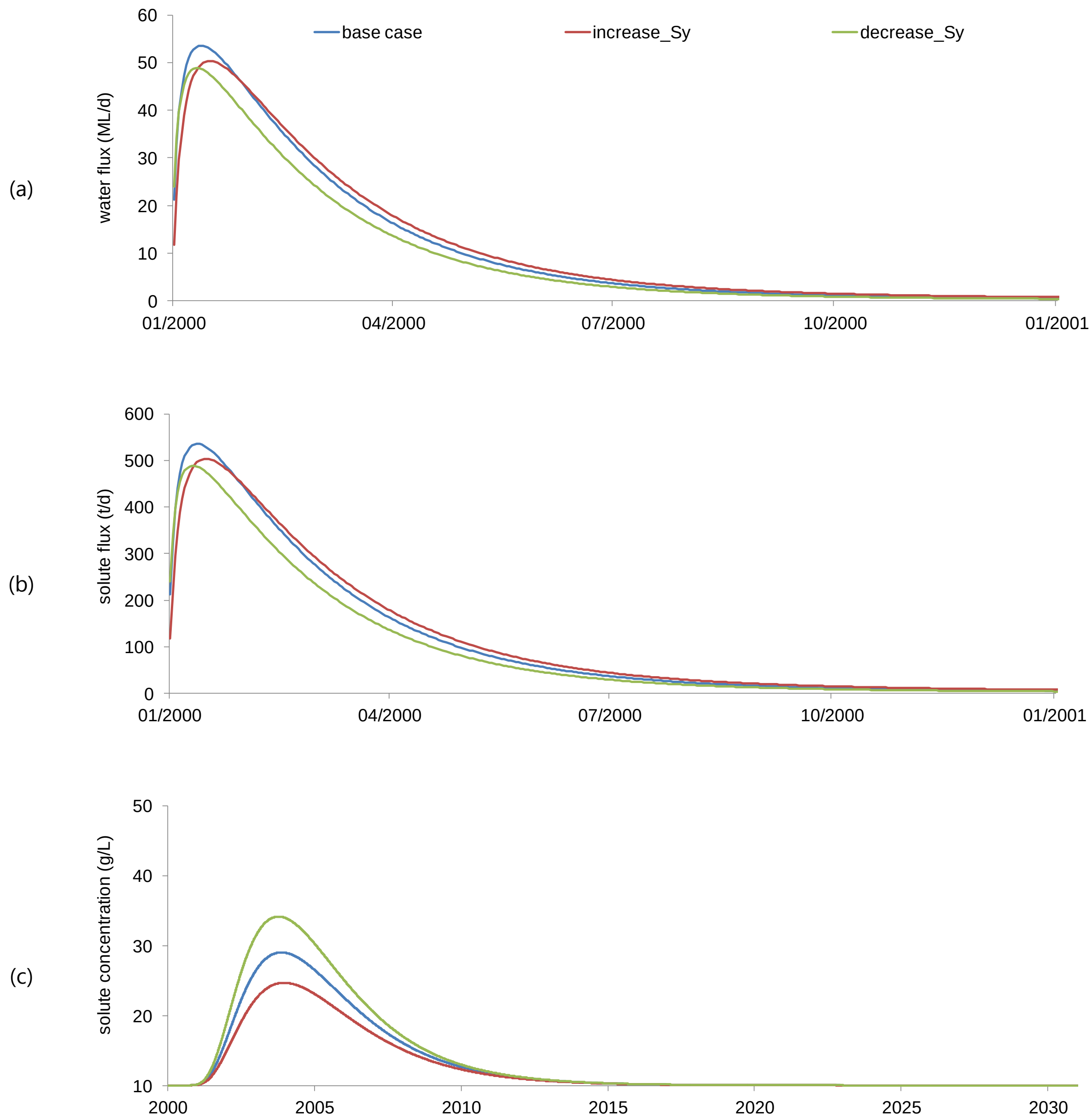


Figure 5-19 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of groundwater discharge to changes in aquifer specific yield. Note that the time scale is compressed for (c).

For the parameter values examined, changes in aquifer specific yield appear to be nonlinearly related to peak water and solute fluxes and the arrival times thereof (Figure 5-19a and b). For example, in comparison to the base case scenario, increases and decreases in specific yield result in reductions in both peak water and solute fluxes. Peak solute concentration is inversely correlated with specific yield, while the timing of this response appears insensitive to variations in specific yield (Figure 5-19c).

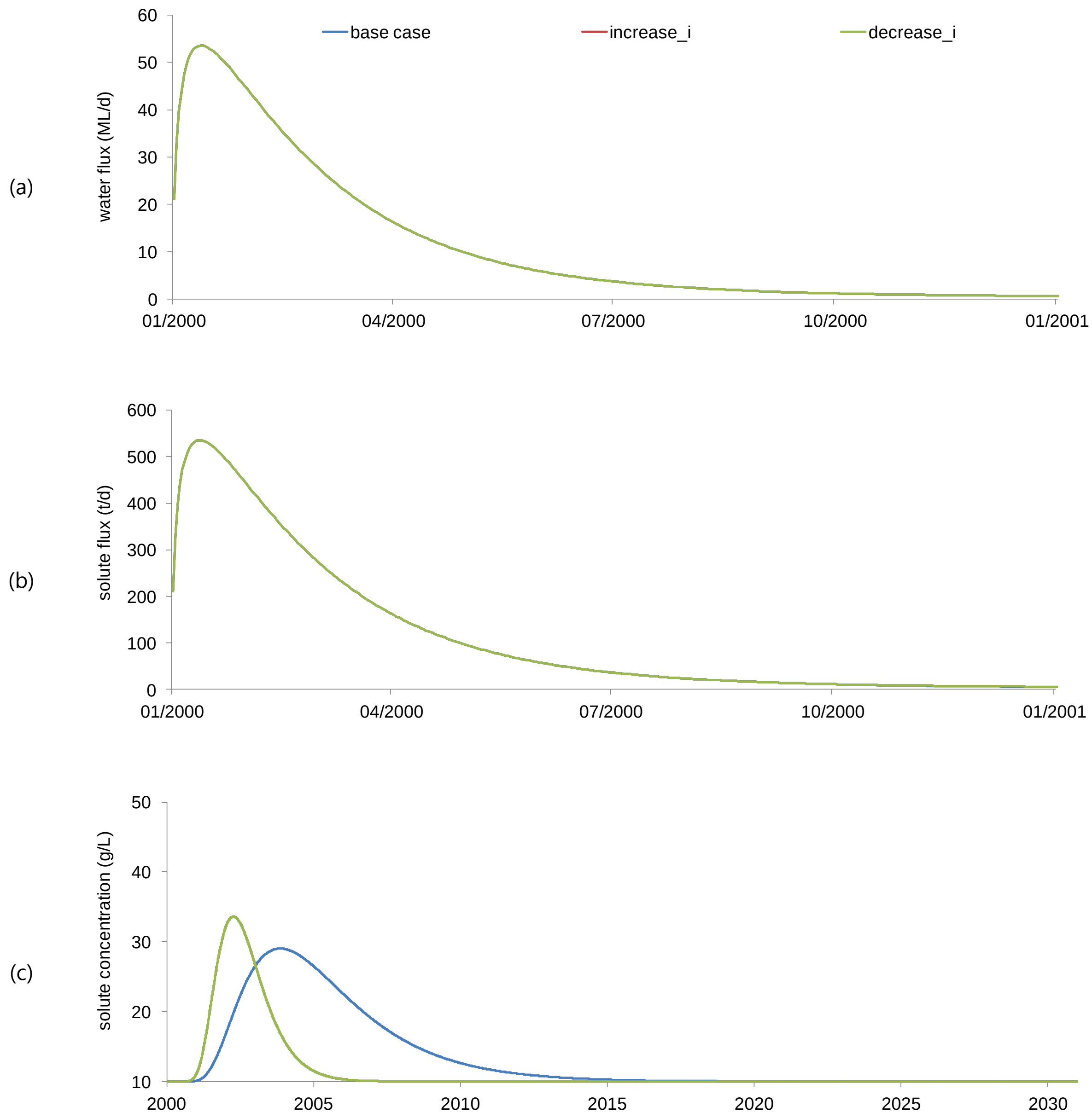


Figure 5-20 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of groundwater discharge to changes in groundwater hydraulic gradient. Note that the time scale is compressed for (c).

Changes in groundwater hydraulic gradient have minimal effect on both water and solute fluxes and the timing of these responses (Figure 5-20a and b). An increase in peak solute concentration and a reduction in the time required to reach peak concentration are observed when the hydraulic gradient is increased. Conversely, inverse trends are observed when the hydraulic gradient is decreased (Figure 5-20c).

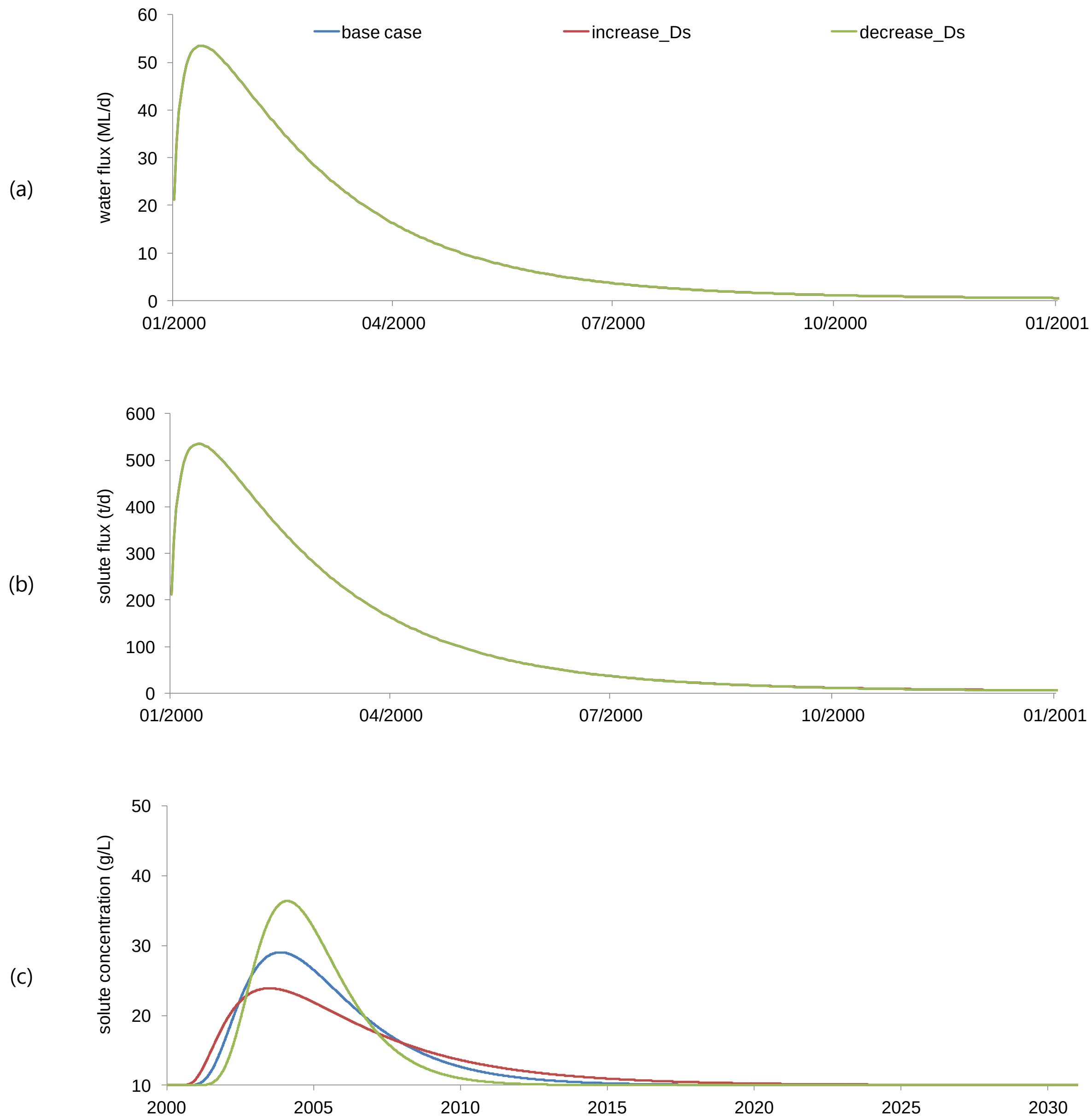


Figure 5-21 Sensitivity of (a) water mass flux, (b) solute mass flux, and (c) solute concentration of groundwater discharge to changes in aquifer solute dispersivity. Note that the time scale is compressed for (c).

Changes in aquifer solute dispersivity have minimal effect on both water and solute fluxes and the timing of these responses (Figure 5-21a and b). Peak solute concentration is inversely correlated with solute dispersivity.

In summary, both water and solute fluxes appear to be insensitive to variations in aquifer porosity, solute dispersivity and hydraulic gradient. The sensitivity of water and solute fluxes to variations in all other parameters tested was also very limited. Peak solute concentration was positively correlated with aquifer hydraulic conductivity and thickness and inversely correlated with porosity, specific yield and solute dispersivity. The timing of peak solute concentration was inversely correlated with aquifer hydraulic conductivity, positively correlated with aquifer porosity, and relatively insensitive to all other parameter variation.

5.3.5 Example 3: Coupled river model

This example aimed to test the operation of the groundwater flowpath in a model that includes surface flows between the river and the wetland. The wetland was initially filled with a small volume of highly saline water to allow the model to develop a residual salt mass in the wetland. After a period of time, overbank flow from the river to the wetland during a high flow event inundates the wetland and subsequently some of this water infiltrates the soil and reaches the groundwater. The infiltrated water and solute were then conveyed back to a downstream section of river via groundwater. The model serves as an example of how the modifications to SOURCE groundwater functionality may be used to represent river–wetland–groundwater interactions.

5.3.5.1 Conceptual model

The model simulates the movement of water and solute from a river to a wetland via a surface channel, then back to the river via groundwater (Figure 5-22). In Source terminology, this was represented using five nodes (i.e. one inflow node, one wetland hydraulic connector node, one storage node, one gauge node, and one confluence node) and numerous links of various types. A temporal extent of 21 years (i.e. 2000-2021) was sufficient for model testing purposes.

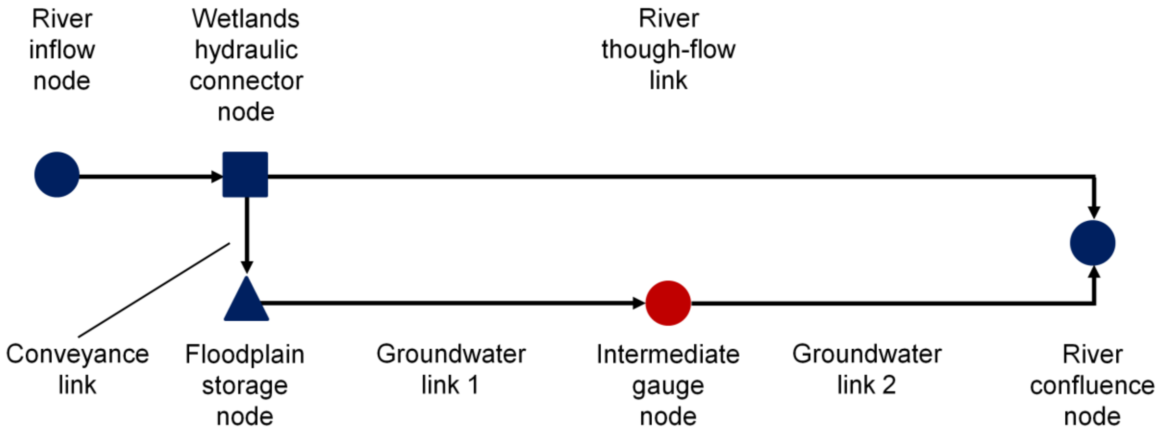


Figure 5-22 Conceptual model used to represent river–wetland–groundwater interactions for testing of the modified version of Source

The river inflow node served as the up-gradient boundary condition and featured a constant flow of 5,000 ML/day except for a three month-long period (June to August) of the second year, during which the inflow increased gradually to a maximum rate of 5,200 ML/day and subsequently declined to return to a rate of 5,000 ML/day. This variation in flow rate was calculated according to a sine function, defined arbitrarily as:

$$Q_{inflow}(t) = 5000 + 200 \sin\left(1.0989\,\pi\,(t - 37043)\right) \quad (5\text{-}15)$$

After this period, river inflows were kept at 5,000 ML/day for the remainder of the 21 year simulation. In practice, a peak flow of 5,200 ML/day on the lower River Murray would not normally be sufficient to instigate overbank flow and wetland inundation. However, for the purposes of the present modelling, this increase in flow serves to initiate flow to, and inundation of, a wetland storage node.

Flow from the river node was directed to a wetland hydraulic connector (WHC) node, which represented the diversion of water to a wetland located adjacent to the river. The volume of diversion was calculated according to the hydraulic gradient between the wetland and the river. Inflows to the WHC node were related to river stage heights through a specified flow versus stage relationship. Essentially, while river flow remained less than 5,200 ML/day, no water was diverted to the wetland. Upon the onset of peak flows, the elevation of water at the WHC node was increased linearly from zero to a maximum of 5 m for a peak flow rate of 5,200 ML/day. The rate at which water was conveyed to the wetland was calculated according to a specified stage versus conveyance flux relationship. For a surface water channel elevation of 5 m, the equivalent rate of water conveyed was specified as 2 $m^{2.5}/s$.

The conveyance link was specified as a bidirectional link. This enabled flows to return to the WHC node from the wetland storage node after the reversal of the hydraulic gradient, which occurred after 24 days of increased river flows.

The return of infiltrated wetland water to the river via groundwater at a location downstream of the WHC node was represented by a confluence node.

5.3.5.2 Parameterisation

The wetland storage node featured an initial volume of 1 ML distributed over an area of 1 km^2. This corresponded to an initial stage height of 1×10^{-3} m. The initial concentration of wetland water was specified as 100,000 mg/l. This combination of limited initial stage height and hypersaline initial concentration represents a nearly dry floodplain in which the evapo-concentration of salt has occurred. Wetland bed conductance was specified as 2000 m^2/day. An evaporation flux from the wetland was also included, at a rate of 5 mm/day.[3]

The groundwater routing link (hereafter "groundwater link 1") was parameterised as in Table 5-4. This link represents flows between the wetland and the river. A constant groundwater concentration is used in the pressure equation, and this example used a value of 100,000 mg/l.

For the storage routing-type link located immediately down-gradient of the previously discussed groundwater link (hereafter "groundwater link 2"), a constant groundwater discharge of 0.1 m^3/day was specified, also with a constant concentration of 100,000 mg/l. Groundwater link 2 represents regional groundwater flow to the river.

All other parameter values and initial conditions specified for the model are summarised in Table 5-4.

[3] The evaporation routine in Source specifies a depth-area relationship, which allows the floodplain to reduce in area as the floodplain shrinks. The total evaporation from the floodplain is therefore not constant, but rather is a function of water depth.

Table 5-4 Model parameter values.

Model element	Parameter	Value	Units
River inflow node	Initial flow rate	5,000	ML/day
	Peak flow rate	5,200	ML/day
	Solute concentration	500	mg/l
WHC* node	maximum average flow	5,200	ML/day
	maximum elevation	5	m
Conveyance link	maximum modified conveyance	2	$m^{2.5}.s^{-1}$
	maximum elevation	5	m
Wetland storage node	wetland area	10^6	m^3
	initial storage volume	1	ML
	bed conductance	2,000	m^2/day
	initial solute concentration	100,000	mg/l
Groundwater link 1	aquifer hydraulic conductivity	1	m/day
	aquifer thickness	20	m
	aquifer porosity	10	%
	aquifer specific yield	0.15	-
	hydraulic gradient	1×10^{-3}	-
	orthogonal distance to discharge location	1,000	m
	solute dispersivity	1	m
	solute concentration	100,000	mg/l
Groundwater link 2	water flux	0.1	m^3/day
	solute concentration	100,000	mg/l

*wetlands hydraulic connector

5.3.5.3 *Results*

Modelling results are presented in terms of water mass fluxes, solute mass fluxes, and solute concentrations. Water mass fluxes reported by the model for the first two years are first presented (Figure 5-23).

Model inflow at the river inflow node (Figure 5-23a) remains constant at 5,000 ML/day until 01/06/2001, at which time it increases gradually to a maximum rate of 5,200 ML/day (on 15/8/2001) before returning to the initial rate of 5,000 ML/day at 01/09/2001. The diversion of water to the wetland storage node (Figure 5-23b) commences at 01/06/2001 and increases to a maximum rate of 170 ML/day at 07/07/2001.

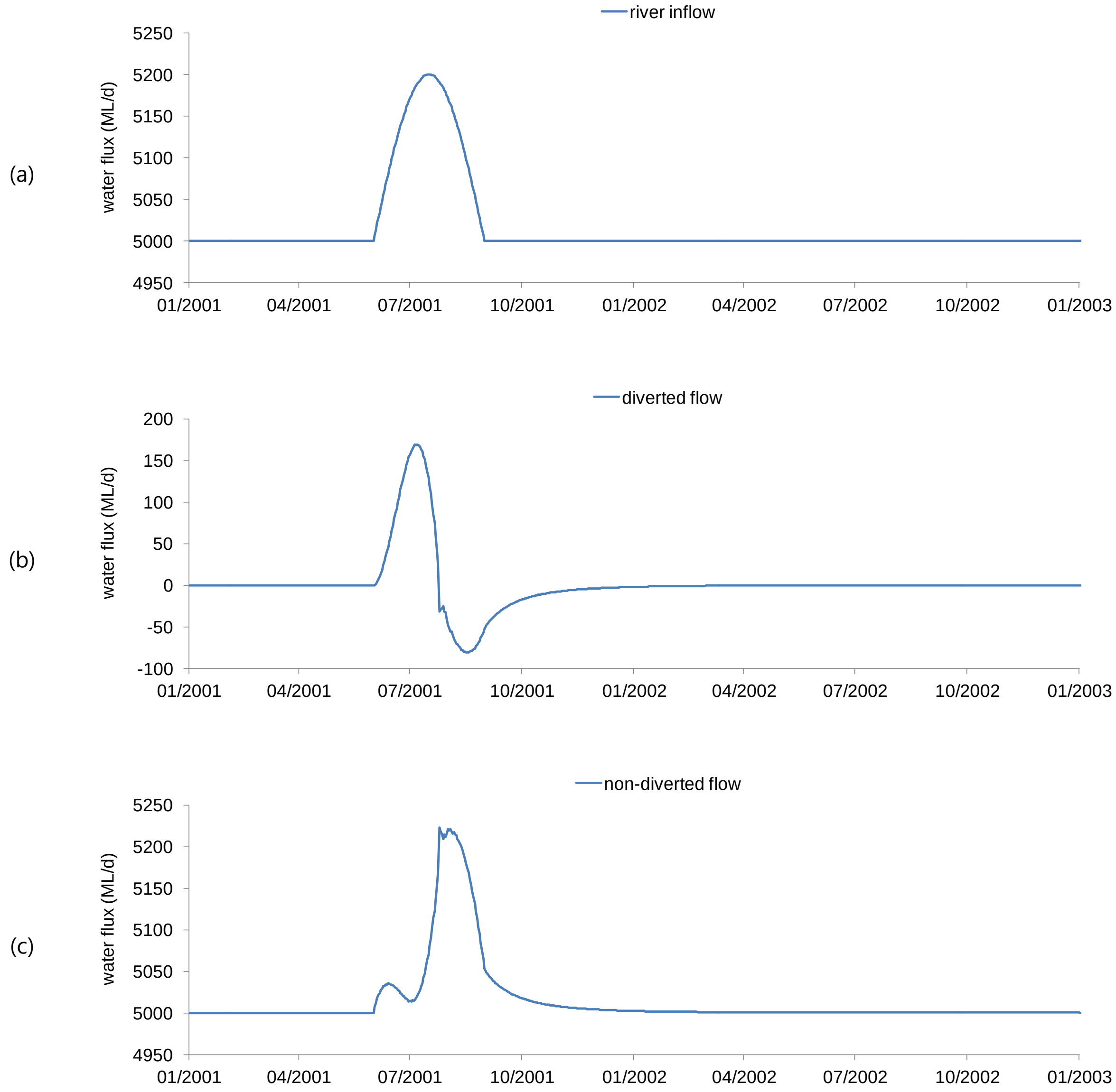


Figure 5-23 Fluxes of water (a) entering the model at the river inflow node, (b) diverted to the wetland storage node via the wetlands hydraulic connector (WHC) node, and (c) flowing downstream from the WHC node.

River flow is diverted to the wetland storage node until 25/07/2001, at which time the hydraulic gradient reverses and water flows back from the wetland to the river. A maximum return flow rate of 81 ML/day occurs on 17/08/2001 and surface return flow persists until 03/08/2001.

Non-diverted flow (Figure 5-23c) increases initially as river inflows increase, then decreases as diverted flows increase. Non-diverted flow subsequently increases, reaching a maximum rate of more than 5,200 ML/day. The additional flow (in excess of river inflow) is derived from return flow from the wetland storage node. Non-diverted flow subsequently decreases, commensurate with the reduction in river inflows.

(a)

(b)

(c)

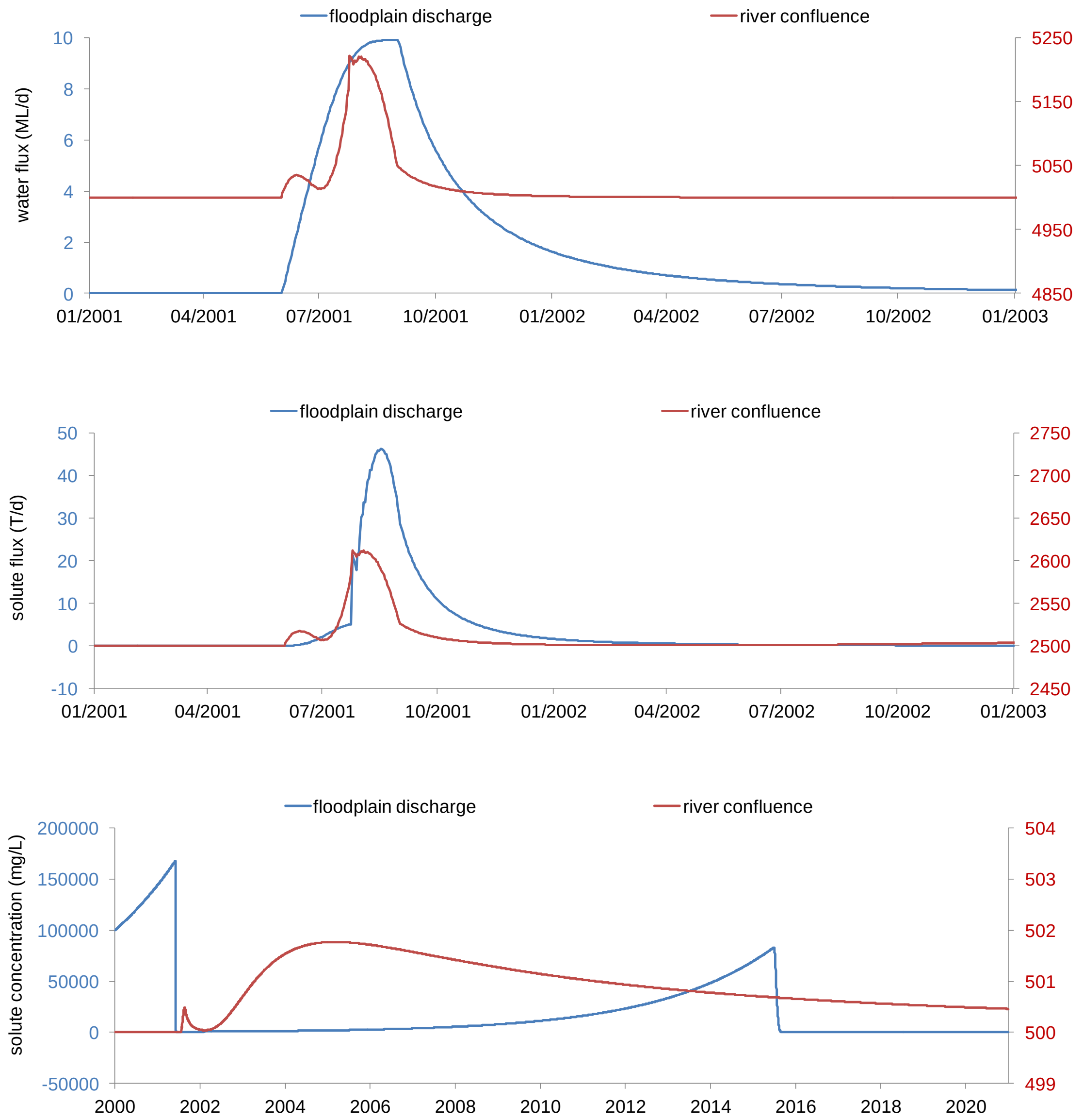


Figure 5-24 (a) Water mass fluxes, (b) solute mass fluxes and (c) solute concentrations of water exiting the wetland (blue) and the groundwater domain (red). Note that the time scale is compressed for (c).

Wetland discharge commences at 2 x 10^{-3} ML/day and gradually declines to ~4 x 10^{-4} ML/day on 02/06/2001, although this is not apparent on the scale used for Figure 5-24a. This represents the infiltration of the initial water volume stored on the wetland. Wetland discharge subsequently increases to a maximum rate of ~9.9 ML/day on 31/09/2001 before declining exponentially to zero over the 16 month-long period following peak river flow. Flow at the river confluence node is almost identical to that observed downstream of the WHC node, which indicates the insignificant contribution of wetland–groundwater discharge to river flow.

The solute mass flux of wetland discharge commences at ~0.2 tonnes/day and gradually declines to ~0.07 tonnes/day prior to the arrival of diverted river flows (Figure 5-24b). The solute mass flux subsequently increases to a maximum rate of ~46 tonnes/day on 31/09/2001 before declining exponentially on 12/09/2007. Solute mass flux dynamics at the river confluence node reflect those of river confluence water fluxes, increasing from 2,500 tonnes/day on 25/07/2001 before declining to ~2,502 tonnes/day by the end of the 21 year-long simulation.

The solute concentration of water exiting the wetland commences at 100,000 mg/l, which is consistent with the specified initial condition (Figure 5-24c). Due to the effect of evapoconcentration, wetland discharge concentrations increase to ~168,000 mg/l at 02/06/2001. The concentration then drops to 500 mg/l during the flooding event. As the wetland dries out, evaporation causes the wetland discharge concentrations to increase to ~83,000 mg/l at 29/06/2015.

In addition to the water flux, solute mass flux and solute concentration of water discharged from the wetland storage node, the mass of solute stored in the wetland was also reported (Figure 5-25). The initial stored mass of ~100 tonnes decreases to ~34 tonnes prior to the arrival of diverted river flows. The stored solute mass then increased to a maximum of ~2,500 tonnes on 24/07/2001 before declining exponentially to zero on 17/07/15. Because Source models leakage as proportional to the water depth, leakage decreases to zero as the water depth decreases to zero. This causes the wetland to drain much more slowly than would usually be the case.

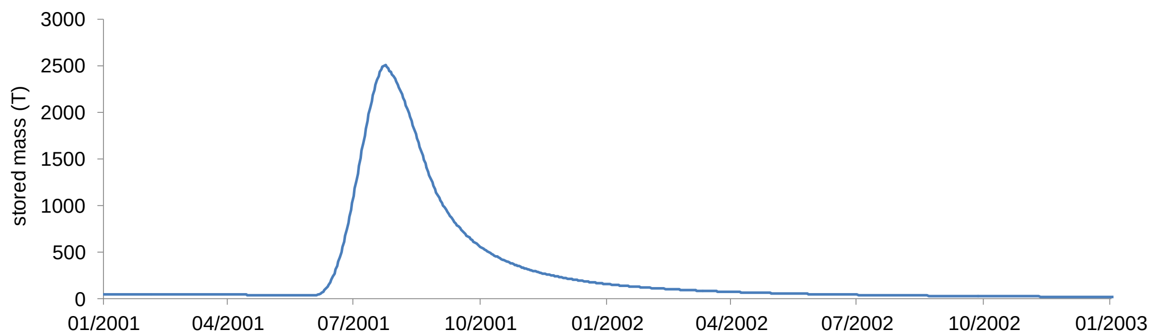


Figure 5-25 Wetland salt mass storage

In this example, a very shallow initial wetland water depth (1 mm) with high concentration (100,000 mg/l) was implemented to create an initial salt mass. Because the evaporation rate (5 mm/day) is much greater than the leakage rate at this water depth, it was expected that the water would evaporate leaving remnant salt that would be mobilised by the next flooding event. However, this did not eventuate because evaporation decreases with water depth. (The evaporation routine in Source essentially assumes that the reduction in water volume represents a reduction in wetland area rather than water depth.) However, as currently implemented, the leakage routine does not operate in this same manner. This means that the wetland inevitably dries out by leakage rather than by evaporation, which means that no salt mass is retained in the wetland. The final output of a zero stored salt mass in this example raises questions about the ability of the current routines in Source to accurately account for the mass of dry salt stored in the surface sediments of a wetland and/or in sediments of the unsaturated zone beneath a wetland.

5.3.6 Summary

The approach for simulating groundwater flow from the wetland to the river functioned efficiently when coded into Source. Sensitivity analysis showed that changes in salt load responded to changes in aquifer parameters as expected from the governing equations. It is important to note, however, that the salt loads calculated from the model represent fluxes from a small area of floodplain (1 km^2) for single inundation events. Although model parameters are within the range of values measured on the floodplain (Woods, 2015a), converting model fluxes to mean fluxes per kilometre of river requires consideration of area and frequency of inundation (per kilometre length of river). This was beyond the scope of the present study, but it is an important next step. However, there are a number of areas where model improvements could be made, and these improvements should precede any comparison with field data. In particular:

- The routines do have the potential to produce negative salt fluxes to the river. This was demonstrated in a model simulation where parameters were deliberately chosen to separate the pressure and transport processes. The problem did not arise in a subsequent model simulation using more realistic model parameters, chosen to be representative of the lower River Murray and its wetlands. However, it is not yet clear how often this problem will arise.

- The current model appears unable to retain salt mass in the wetland if it dries out between flood events. It appears that this occurs because the evaporation routine specifies a bathymetric function to reduce evaporation as wetland volume decreases (simulating a reduction in wetland area). However, this bathymetry function is not currently used by the leakage routine. This model inconsistency should be relatively straightforward to fix. However, it currently means that the wetland tends to dry by leakage rather than by evaporation, and so all of the salt is removed with the leakage.

- The model assumes that leakage is proportional to water depth. In reality, however, gravity drainage should still occur when the water depth is zero. By not including this process, the wetland dries more slowly than should be the case. Whether this discrepancy is significant for modelling salt fluxes to the river has yet to be determined.

5.4 Floodplain Groundwater Velocity and Travel Times

5.4.1 Outline

One of the critical parameters used by the above model is the regional groundwater velocity (which is calculated from the regional groundwater gradient, the aquifer hydraulic conductivity and the porosity). The model assumes that this velocity is constant, whereas in reality it is likely to change over time due to infiltration events on the floodplain. Here, we compare the timing of salt loads to the river calculated using a constant velocity with those calculated by a model that allows the groundwater velocity to be calculated from the wetland leakage. The approach is directly analogous to flow in a pipe, with the wetland forming one end of the pipe, and discharge to the river at the other end. Water only moves through the pipe when the wetland is leaking. (At other times the velocity is zero.) Because the volume of the pipe is fixed, when water enters the upstream end of the pipe through the base of the wetland, water immediately moves out the other end of the pipe into the river. This represents the pressure effect. Of course, the model represents the pressure effect as instantaneous, whereas in reality it would be somewhat delayed. A third model includes vegetation extraction along the flow path. The mechanics of the three models is described below.

5.4.2 Formulation

5.4.2.1 Surface Water Module

The main input data for all the models are river stage heights over time. Flow between the river and wetland is then controlled by a sill height. When water level in the river (or wetland) exceeds the sill height, then flow between the two water bodies is calculated using the Manning Equation. (Water flows from the river to the wetland , when the river stage exceeds the sill height and exceeds the water level in the wetland . Water flows back from the wetland to the river when the water level in the wetland exceeds the sill height and exceeds the river stage.) The wetland area is constant, so that changes in the water depth in the wetland are given by

$$\frac{\mathrm{d}h}{\mathrm{d}t} = \frac{Q_{in} - Q_{out}}{A} - E - q \qquad (5\text{-}16)$$

where h is the wetland water depth, Q_{in} is surface flow from the river to the wetland , Q_{out} is surface flow from the wetland to the river, E is the evaporation rate from the wetland , q is the wetland leakage rate, A is wetland area and t is time. The river stage is prescribed in the model input file and does not change due to water flow to the wetland. Similarly, changes in the wetland salt store are given by:

$$\frac{\mathrm{d}M}{\mathrm{d}t} = \frac{Q_{in}c_{riv} - Q_{out}c}{A} - qc \qquad (5\text{-}17)$$

where M is the salt mass in the wetland , c is the solute concentration in the wetland , c_{riv} is the concentration in the river (which is assumed to be constant in time). The salt concentration in the wetland is simply calculated as the mass divided by the water depth ($c=M/h$), except that it is constrained by a maximum concentration (equivalent to the saturation point). Thus when M/h exceeds the saturation point, the salt is assumed to be only partly dissolved, and partly as precipitated crystals. The model restricts the fluxes to the dissolved part.

The leakage rate beneath the wetland is assumed to be proportional to the water depth, and is given by:

$$q = \frac{K_c(h + h_c)}{h_c} \qquad (5\text{-}18)$$

where h is the depth of water in the wetland, h_c is the thickness of a low permeability layer at the base of the wetland, and K_c is the saturated hydraulic conductivity of this layer.[4]

5.4.2.2 Groundwater Module

Three different models were constructed, all based on slightly different process conceptualisations. The different conceptualisations lead to different travel times for salt between the wetland and the river. In each case, the salt mass flux to the river is calculated as

$$m_r(t)\frac{w}{A} = wq(t - t_s)C_L(t) \qquad (5\text{-}19)$$

where t_s is the salt travel time between the wetland and the river, and C_L is the salt concentration of wetland leakage, and t is time. Note, however, that the travel time is not necessarily a constant value, but can change over time. The left hand side of Equation 5-19 is the salt flux to the river per unit length of river that parallels the wetland.

The three models are summarised as:

[4] This formulation is slightly different to that used by the Source model. (Compare Equations 5-11 and 5-18.) The Source model assumes that leakage is proportional to water depth, whereas the Fortran model assumes that leakage is proportional to ($h+h_c$). The latter approach is more correct, because if the floodplain soil is saturated, then the floodplain will still drain even though there may be no surface water (Brunner *et al.*, 2009).

1. **Constant velocity model**. This model moves infiltration back to the river at a constant velocity, which is specified. This is the approach used for the transport component of the SOURCE model. The travel time (t_s) of salt between the wetland and the river is therefore given by

$$t_s = \frac{x}{v} \tag{5-20}$$

where *x* is the distance between the river and the wetland (see Figure 5-4) and *v* is the velocity. The salt flux to the river is then simply calculated using Equation 5-19. The model does not allow the leakage of water beneath the wetland to affect this velocity. The model is therefore perhaps most appropriate for a system with a large regional groundwater gradient, where the mean leakage rate beneath the wetland is small relative to the regional flow rate. This approach does not include a pressure effect, but this may be unimportant if there is a strong regional hydraulic gradient and the volumes of recharge are small. Of course, the problem with implementing this model is that the groundwater velocity must be known.

2. **Pipe model**. This model allows the velocity of water to change in time, based on changes in the wetland leakage rate. Each parcel of infiltration is essentially pushed by subsequent infiltration. When the wetland is dry and there is no infiltration, the groundwater velocity is zero. The water velocity is a function of the infiltration rate beneath the wetland, given by:

$$v(t) = \frac{q(t)w}{b\theta} \tag{5-21}$$

where *w* is the width of the wetland and *b* is the aquifer thickness (Figure 5-26). The travel time (*t'*) between infiltration and discharge for a particle leaking beneath the wetland at time t_0 can be calculated from:

$$xb\theta = w \int_{t_0}^{t_0+t'} v(t)dt \tag{5-22}$$

where *x* is the distance between the wetland and the river and *v(t)* is the water velocity along the flowpath.

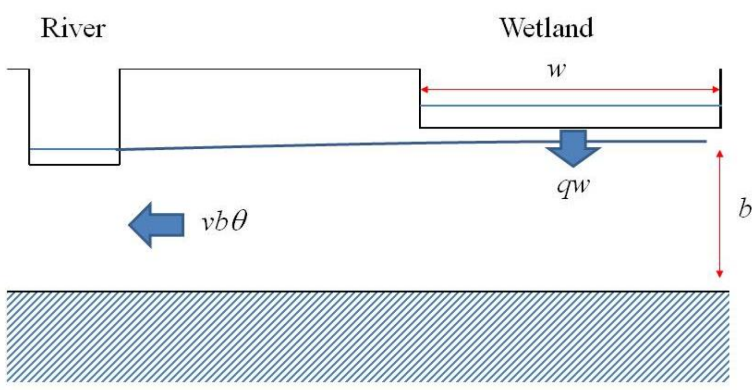


Figure 5-26 Schematic representation of groundwater flow from a wetland to a river. The wetland leakage flux is *qw*, and the flux to the river is *vbθ*. Both quantities have units of L^2/T, and represent fluxes per metre length of river.

3. **Pipe Model with Vegetation Extraction**. This model moves water towards the river as the single pipe model, but also allows water to be extracted by vegetation along the flowpath. This reduces the groundwater velocity. This process also concentrates salt, however the total salt flux does not change – the discharge is just delayed. The travel time (*t'*) between infiltration and discharge for a particle leaking beneath the wetland at time t_0 can be calculated as:

$$aH\theta = w \int_{t_0}^{t_0+t'} v(t)dt - E_t a t_t \, dt \tag{5-23}$$

where E_t is the evapotranspiration rate along the flowpath. (The model actually uses a slightly different approach, in which *x* is replaced by *x-w*/2 if the wetland contains water and *x*+*w*/2 if the wetland is dry. This allows for evapotranspiration to also occur beneath a dry wetland.)

5.4.3 Example

The model was run using historic water level data from Lock 5 as the input. Data from 1 January 1936 till 1 July 2013 was used, during which time the river level varied between 12.55 and 19.46 m. The data was duplicated to form a sequence of 155 years of daily river stage data for the simulations. Duplication of the data was necessary because of the long travel times along the groundwater flowpath between the wetland and the river. Other model parameters are given in
Table 5-5.

Table 5-5 List of groundwater parameters for different models

Symbol	Description	Value	Units
A	Area of wetland	100,000	m^2
b	Aquifer thickness	100	m
c_{riv}	Concentration in river	500	mg/L
c_{max}	Maximum permissible concentration	200,000	mg/L
E	Wetland evaporation rate	0.005	m/day
[1]*Et*	Evapotranspiration rate	0.001	m/day
hc	Thickness of wetland bed	1	m
Kc	Hydraulic conductivity of wetland bed	0.02	m/day
[2]*v*	Groundwater velocity	0.05	m/day
w	Width of wetland	100	m
x	Distance of wetland from the river	250	m
θ	Aquifer porosity	0.4	-

[1] Vegetation model only
[2] Constant velocity model only

Figure 5-27 shows the river level, wetland level, flow rate between the river and wetland and leakage rate beneath the wetland over half of the simulation period. For these simulations the base of the river was assumed to be at 15 m, and the sill height was 16 m. As can be seen flows from the river to the wetland commences only when the river level exceeds the sill height of 16 m. Return flow from the wetland to the river (shown as negative flows on Figure 5-27c) occur when the river level drops below the sill height. Leakage beneath the wetland occurs whenever the wetland contains water (level above 15 m in this simulation). The leakage rate is proportional to the depth of water in the wetland.

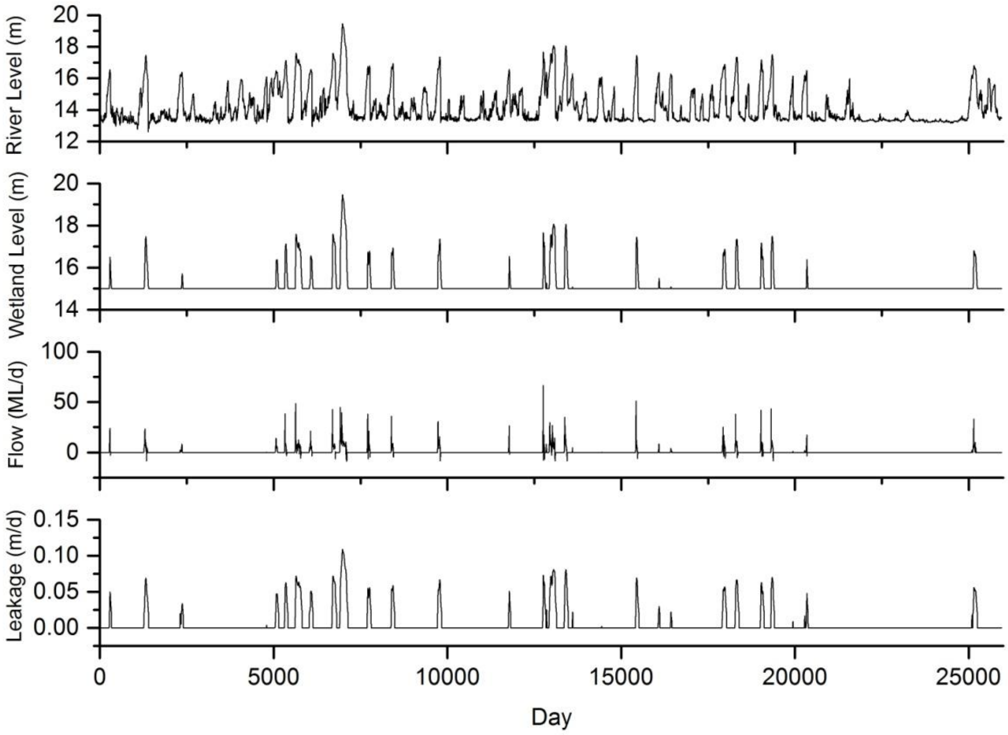
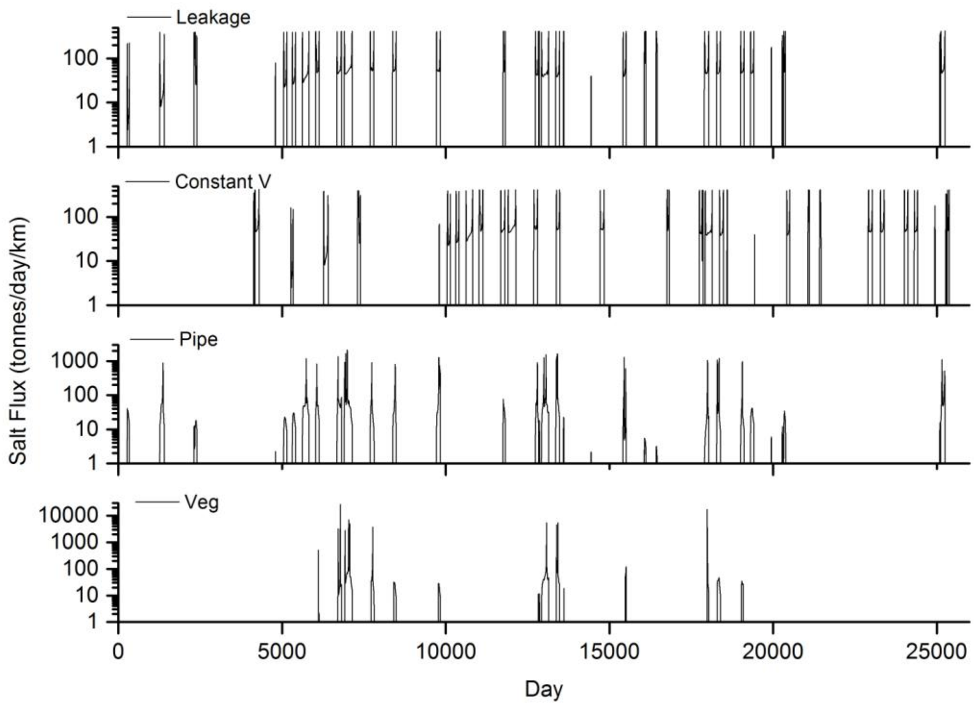


Figure 5-27 River and wetland levels, flow rate between the river and the wetland and leakage rate beneath the wetland for a sill height of 16 m. The leakage rates form the input flux for the groundwater flow module in each of the different models.

The surface water fluxes and leakage rate are the same for all three models. Figure 5-28 and Figure 5-29 show wetland leakage salt flux (also the same for all three models) and the salt flux to the river for the three different models. (Figure 5-28 uses a logarithmic scale for salt flux, whereas Figure 5-29 uses a linear scale.) Note that wetland leakage salt flux for each event is initially high, and then drops, but increases again as the wetland dries out. This is shown more clearly in Figure 5-30, which highlights the salt leakage from the flood event between 5600 and 5850 days. The initial high rate of salt infiltration is due to salt which was stored within the wetland prior to the flood event. As flood waters enter the wetland, they bring in relatively fresh water, and so the salt flux decreases (even though the rate of water leakage may increase). (The water leakage rate peaks at approximately 5650 days in this case.) After 5780 days, both the river and wetland have dropped below the sill height. The subsequent increase in flux is due to the increase in salinity in the wetland due to evaporation.

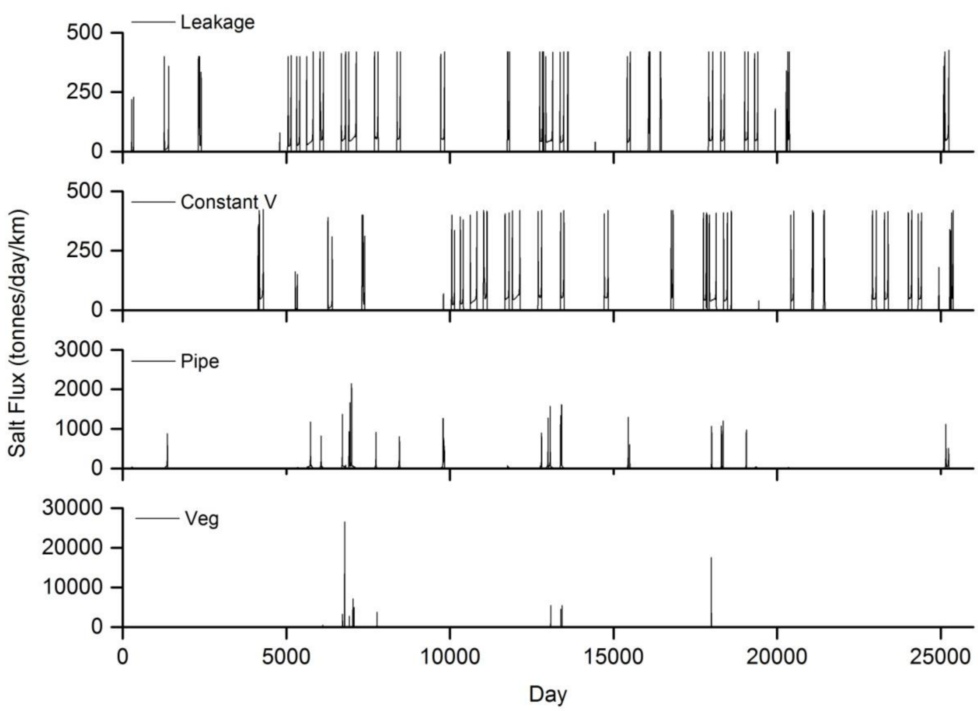
The maximum salt flux in wetland leakage over the simulation period is approximately 427 tonnes/day/km, and these high fluxes tend to occur when the concentration is high and the leakage rate is relatively low. For the constant velocity model, a constant velocity of 0.05 m/day was specified. The salt flux to the river is therefore simply equal to the salt leakage flux, delayed by a travel time of 5,000 days. (The wetland distance is 250 m, and



hence 250 m / 0.05 m/day = 5,000 days.) For the single pipe model, the maximum salt flux to the river is 2,140 tonnes/day/km. The higher salt flux occurs because groundwater velocity is high during leakage events, and so more saline water is displaced to the river. For the vegetation model, the saline groundwater becomes concentrated by evapotranspiration, and so this effect is more pronounced, with a maximum salt flux to the river of 26,600 tonnes/day/km.

For the pipe model, each wetland leakage event is accompanied by a salt discharge to the river. This is the pressure effect of leakage, which is instantaneous in this model. In fact, the water leakage rate will be equal to the water flux to the river. However, the salt flux to the river is different to the salt flux in leakage, because the water may have a different concentration.

Figure 5-28 Salt flux leaking beneath the wetland, and salt flux to the river under three different



groundwater flowpath conceptualisations. Note that y-axis scales are logarithmic.

Figure 5-29 Salt flux leaking beneath the wetland , and salt flux to the river under three different groundwater flowpath conceptualisations. This figure contains the same data as Figure 26, except that salt flux is shown using a linear scale. Because of this, some of the smaller salt fluxes are not visible on this figure.

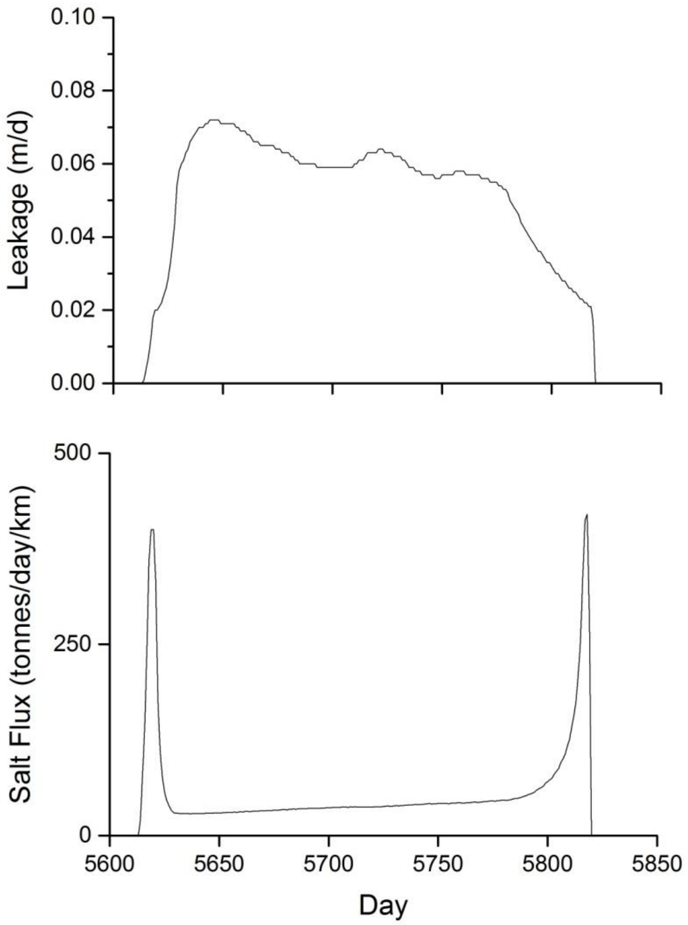


Figure 5-30 Water and salt flux beneath the wetland for a single filling and draining event

Figure 5-31 indicates travel times to the river for salt infiltrating in two particular events: one centred around Day 5,070 and a second centred on Day 8,400. As discussed above, travel times for the constant velocity model are 5,000 days in each case, and so this salt arrives at the river in Days 10,070 and 13,400 respectively. For the pipe model, travel times are much longer. This is because there is no flow towards the river when the wetland is dry. The travel time for the first event is 13,230 days, and hence it reaches the river on Day 18,300. The travel time for the second event is 23,630 days, and so it reaches the river on Day 32,030. The first event travels more quickly to the river, because it is followed by a number of other infiltration events, which push it. In contrast the second even occurs before a dry period, and so it moves more slowly. The vegetation model has even longer travel times because extraction of water by the vegetation slows down the water velocity.

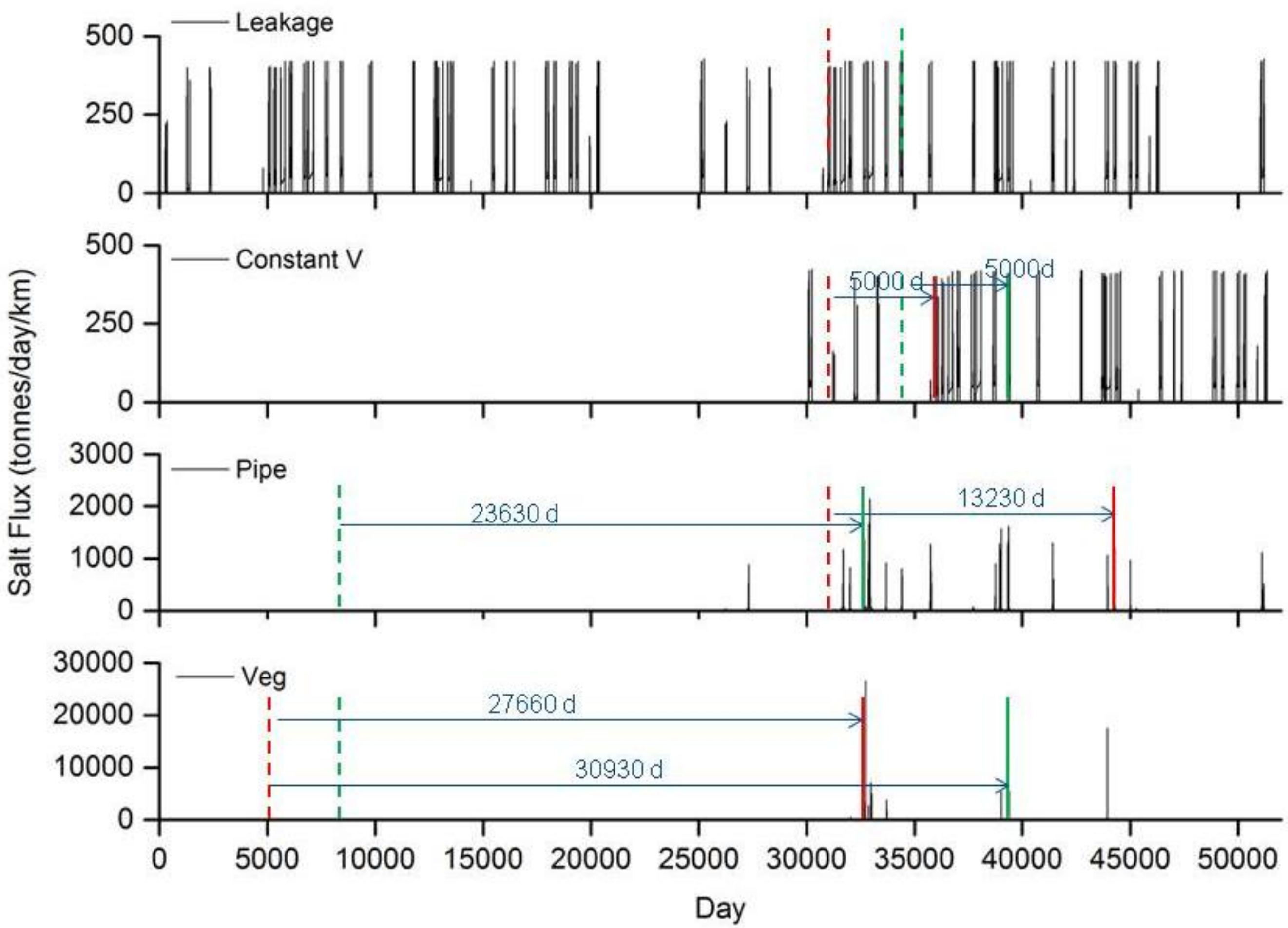


Figure 5-31 Salt flux leaking beneath the wetland, and salt flux to the river under four different groundwater flowpath conceptualisations. The travel time of leakage is shown for each of the models, for two particular leakage events. Salt flux to the river is not shown for the first 25,000 days, due to model warm-up (i.e. equilibration of initial conditions).

5.4.4 Summary

Three different methods of calculating groundwater velocity between a wetland and river were compared. The first conceptualisation used a constant velocity and resulted in salt fluxes to the river that were unrelated to river flow rates. The second approach, whereby the groundwater velocity was related to the wetland leakage rate, resulted in salt fluxes to the river that only occur at times of wetland leakage. This model produced larger peak salt loads to the river than are produced by the constant velocity model. The model with vegetation extraction of groundwater has the largest peak salt loads and longest groundwater travel times.

The use of a constant velocity for the movement of salt between the wetland and river is likely to be realistic if the regional groundwater velocity is high relative to the wetland leakage rate. However, whether this condition is met across all of the River Murray floodplain areas is unclear. The comparison in this section, shows that salt fluxes and travel times are highly dependent upon the method used to determine the groundwater velocity. Further analysis of how groundwater velocity changes with time within river floodplain environments would therefore assist our understanding of salt transport between wetlands and the river.

5.5 Discussion and conclusions

The Source model (Welsh *et al.*, 2012) has been widely promoted as a tool for simulating river flows and dissolved salt concentrations. The model also includes some limited groundwater functionality, with the capacity to simulate changes in groundwater flow to the river due to irrigation leakage or groundwater pumping. The current project has developed a simple approach for including salt transport on river floodplains. Surface water exchange between the river and inundated areas of the floodplain are simulated using existing surface water routines within the model. The new routines allow flow of water and salt back to the river through groundwater pathways.

There are two important components to this salt movement. The first is related to the pressure response created by the groundwater mound caused by leakage of floodwaters into the regional aquifer. The salt flux induced by this pressure response is unrelated to the salinity of the leakage water. The second component is the transport of the leakage water back to the river. This transport occurs more slowly, but the salt flux is determined by the concentration of the leakage water. An approach for simulating both the pressure and transport processes between a wetland and the river was developed and incorporated into a version of the Source model. The model represents areas of floodplain inundation using wetlands. These wetlands fill from overbank flows from the river, and dry by surface water return flows, evaporation and leakage. Although each wetland is represented using simple, average parameter values (e.g., wetland bed thickness and hydraulic conductivity), more complex patterns of inundation could be represented using multiple wetlands.

A number of simple simulations were performed using the model to test that the model equations were coded correctly and that the calculated water and salt fluxes to the river were consistent with the conceptual model. However, due to time constraints within the project, only a relatively small number parameter values were used, and hence further testing is desirable. Based on the testing carried out to date, a number of observations and areas where the model could be further improved were noted:

1. The approach has the potential to calculate negative groundwater discharge concentrations under certain conditions. This was not encountered during the initial simulations, but was demonstrated in a specific simulation where parameters were deliberately chosen to highlight the problem. It should be pointed out that the problem did not arise in any model simulations using more realistic model parameters, chosen to be representative of the lower River Murray and its floodplains. However, it is not yet clear how often this problem will arise, and further analysis is desirable.

2. The current model routines assume a constant wetland area for calculation of leakage fluxes, but the evaporation routine used by Source assumes that wetland area varies with floodplain stage. In principle, this problem is straightforward to fix, but this has not occurred due to time constraints in the current project. As it stands, this discrepancy between the two routines means that floodplains inevitably dry out by leakage rather than by evaporation. As a consequence, the floodplain salt mass is completely lost (as leakage) when the floodplain dries out. This is not realistic, and will probably mean that salt loads to the river will be underestimated.

3. The approach uses a constant velocity for transport of salt between the wetland and the river. Through a separate set of simulations using the Fortran code, we have examined the effect of this assumption on travel times and salt fluxes to the river. Clearly, how the groundwater velocity between the wetland and the river is calculated affects the timing of salt loads, but it also affects the magnitudes of salt loads to the river. There is a need to further examine groundwater velocities within floodplain environments, and how these may change with time. This could be undertaken both by examining results from a 3D numerical groundwater flow and solute transport model of the floodplain, but also from field studies designed to measure groundwater flow rates.

4. The model calculates flow to the river independent of the river level. It thus permits groundwater discharge to the river during high river stages. In reality, flow of river water into the aquifer will occur at these times, and groundwater discharge will not occur until river stage recedes. The impact of this on the simulation results has not been specifically examined. The GSWIT module in Source models the water fluxes to and from the river that result from stage height variations. The interaction between this module and the routines developed in this project have not yet been investigated. This may partially address this issue.

A number of the above issues are perhaps best examined by comparing Source model runs with parallel runs using a numerical code such as MODFLOW. This work is currently being planned.

6 Conclusions and recommendations

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6.1 Conclusions

South Australia is seeking to improve understanding of the short-term movement of water and salt within the floodplain landscape. Schedule B to the Murray-Darling Agreement establishes a requirement to identify, assess, report, monitor and review management actions which cause a significant long term increase in the salinity of the River Murray at Morgan: in SA this has been addressed using the SA Salinity Register models and Source model. However, the Murray Darling Basin Plan expands obligations to manage and report on short-term salinity and water quality targets for the River Murray. At the same time, there have been changes in the management of River Murray floodplains and a variety of techniques are being trialled to improve the health and robustness of floodplain ecosystems. To meet these needs, this project has considered how to represent floodplain processes to inform floodplain and river salinity management. The outputs will enable the progressive development of models of the SA River Murray to simulate the impact of environmental actions on groundwater flow and salinity, including exchanges with the River Murray and freshening of floodplain aquifers. They will also inform models of other regions where surface water – groundwater interactions are important.

This project has examined the data required to effectively understand and manage floodplain salt loads, and has also examined two different modelling approaches for quantifying the potential effects on river and floodplain salinity. The approaches were selected for this project as they would immediately augment existing models and capabilities at the SA Department of the Environment Water and Natural Resources (DEWNR). The first approach uses a numerical groundwater flow and transport model (MODFLOW with MT3DMHS). The second approach develops simple routines to estimate salt loads to the river due to leakage of water beneath inundated areas of the floodplain. These routines are suitable for incorporation into hydrologic river models, and have been included in a version of the Source model. Some limited testing of the routines has also taken place.

These models represent only two of a large possible choice of modelling approaches. Although MODFLOW is broadly representative of a range of numerical groundwater models, and Source is representative of a range of similar river models, it is instructive to consider when we should use these models, and when other models would be preferred. The choice of model will be influenced by the model aim, what we believe are the most important processes, and the accuracy with which the different models describe these processes. It will also be influenced by the time required to construct and run the model, and the ability to carry out the required number of simulations in the required time period. It will also be influenced by the familiarity of the modeller (and the modeller's institution) with different numerical codes. This is a complex discussion, and but the following points are noted:

- Current approaches to modelling the lower River Murray and SA floodplains often involve the simulation of a single model simulating a single hydrological domain, without comparison to models of other domains (e.g. a groundwater model is developed without comparison to surface water models of the same area). For a region with considerable interaction between groundwater and surface water, a multi model approach is preferred, where separate models of different hydrological domains are co-developed to be consistent with each other. Another option is to use a fully integrated model which simulates multiple domains simultaneously.
- MODFLOW is suitable where a model's primary aim is to estimate groundwater fluxes (or salt fluxes, if used in conjunction with the solute transport code MT3DMS) over large spatial and temporal scales. It simulates groundwater flow and transport in a spatially distributed manner throughout the floodplain aquifer, and allows spatial variation in parameters to be included where these are known. Its modular design and public domain software status has made it the de facto standard code for most groundwater models. It accounts for the effects of surface water hydraulics through relatively simple sink/source fluxes, without specifically simulating flow in the river, and so results are expressed in terms of salt loads rather than river salinities.
- A variable-density coupled groundwater flow-and-transport model, such as SEAWAT or SUTRA, is necessary for the accurate simulation of groundwater features such as freshwater lenses, where there are significant variations in groundwater salinity.

- The Source model is focussed on predicting impacts on the river, rather than impacts on the floodplain itself. It is therefore not suitable when most interest is on areas of inundation, salinities within the floodplain, and floodplain health. It is intended to predict impacts of the floodplain on the river, and not impacts of the river on the floodplain. The model has fast run times, and so it best suited to large-scale simulations (e.g., whole of River Murray in South Australia).
- Two-dimensional hydrodynamic models have also been widely used for simulating floodplain inundation. These would be preferred where it is important to accurately simulate the spatial and temporal patterns of floodplain inundation.
- Fully-integrated models such as HydroGeoSphere may be suitable for applications involving small or coarsely-discretised domains where simulation runtimes remain manageable. The advantage of fully-integrated models is that they explicitly simulate multiple hydrological domains, including groundwater flow, unsaturated zone flow, and flow and solute transport in the river. Simulation of the river allows salt flows to the river to alter the salinity of the river, and hence change the subsequent concentration of river water flowing into the aquifer. However, HydroGeoSphere does not simulate evapoconcentration, limiting its application for saline floodplains. MIKE SHE is another fully-integrated option worth investigating.

It is recommended that where multiple models are developed of a region of interest, that consistency is maintained between the models. For example, a MODFLOW model of groundwater flow may use the results of a 2D hydrodynamic model and an unsaturated zone model as inputs, while the flux between the river and the floodplain aquifer could be compared with the results of a Source model. Discrepancies would identify gaps in conceptual understanding and in data.

6.2 Recommendations

The companion report, *Modelling salt dynamics on the River Murray floodplain in South Australia: Conceptual model, data review and salinity risk approaches* (Woods, 2015a), provides recommendations regarding the conceptualisation of saline floodplain dynamics, including some tasks which require modelling. The recommendations address: (i) improving conceptual understanding using existing datasets and models, (ii) how to address key conceptual gaps, and (iii) the development of a Salinity Risk Framework. The recommendations provided here are in addition to those of the companion report.

6.2.1 Data and monitoring to support conceptual understanding and numerical models

Where detailed modelling is required of an area's floodplain salinity dynamics, the modelling should be supported by a monitoring program. The scope of the program will depend on the aims of the project, budget and time. Section 2.5.1 provides an exhaustive list of monitoring tasks which would inform the modelling. For many of these items, there will be existing data and infrastructure which should be reviewed before a monitoring plan is developed. If a preliminary model is available, it can be used to identify which parameters the model results are most sensitive to, so that these can be prioritised.

This study has also highlighted the importance of understanding how groundwater heads and flow velocities change with time across the floodplain. The magnitude and timing of salt flows to the river arising from groundwater recharge beneath inundated wetlands is highly dependent on the groundwater velocity. We currently do not have a good understanding of the extent to which the groundwater velocity across the floodplain changes due to floodplain recharge events.

6.2.2 MODFLOW modelling of saline floodplains

The table below provides preliminary recommendations for MODFLOW modelling of salt dynamics in SA River Murray floodplains. The importance of the recommendations will depend on the model aim. The recommendations are based on the conceptualisations and parameters adopted for testing in Chapter 4, so the recommendations may not be as applicable in regions with significantly different hydrogeology. These recommendations are currently guiding the development of DEWNR's Pike Floodplain numerical model as part of the SARFIIP Salinity Management Measures.

Table 6-1 Preliminary recommendations for MODFLOW modelling of SA River Murray floodplains

Topic	Recommendation	Reason
Software selection	Use MODFLOW2005 with the NWT solver and UPW package.	This simulates the wetting and drying of the Coonambidgal Formation more consistently than earlier versions of MODFLOW. MODFLOW-Surfact may also be used, but is not compatible with the most-commonly used solute transport code for MODFLOW, MT3DMS. See Section 4.2.8.2.
Solute transport	Simulate explicitly, e.g. using MT3DMS, rather than estimating salt transport in groundwater by multiplying flux by a constant salinity.	Freshening of groundwater by river water lowers groundwater salinity near the river, so a constant-salinity calculation may overestimate the salt flux in a dynamic river system. See Section 4.4.3.2.
Stress period length	If the key model output is potentiometric head or ET flux, use monthly stress periods. If the key model output is groundwater flux/salt to river, use stress periods that change in length depending on the gradient of the river stage, i.e. shorter stress periods during periods of high flow. Woods (2015b) Appendix A includes a Python code for selecting stress periods based on river level.	Numerical experiments show that using yearly averages of river stage and/or ET does not capture seasonal and climactic variations in potentiometric head. Monthly and sub-monthly stress periods do capture this behavior. However, flux between river and groundwater is very sensitive to stress period length, so sub-monthly, "adaptive" stress periods are recommended. See Sections 4.4.1 and 4.3.2.
Evapotranspiration	Develop an understanding of how evapotranspiration relates to depth and choose an appropriate representation of this, which may include using a piecewise linear ET function.	The floodplain water balance and river condition (gaining or losing) are very sensitive to the shape and parameters of the ET function. At this stage, a lack of data precludes more specific advice on ET simulation. See Section 4.3.2.
Topography	Use as accurate a representation of surface topography as possible, within the limits of the available data. Consider also the computational time required for simulating fine grids.	ET is sensitive to the depth to water and hence to the topography. As noted above, the floodplain water balance and river condition are very sensitive to ET. See Section 4.3.2.
Flood inundation	Identify areas of likely higher inundation recharge by examining AEM, vegetation and soil data.	Inundation recharge rates are rarely measured in the field but may be inferred. See Section 4.4.3.
Overall	Incorporate outcomes of work done to address conceptual gaps.	The details of some key floodplain processes are still poorly known or quantified. Advances in knowledge should inform models, e.g. revised inundation recharge rates from unsaturated zone investigations, and constraining the model to match AET observations.

6.2.3 Further modelling

Floodplain salinity dynamics are extremely complicated. This initial study conducted only the most urgent investigations. Table 6-2 provides suggestions for further scenarios that could be investigated using the MODFLOW model discussed in Chapter 4. Table 6-3 gives recommendations for groundwater modeling that involve other models and/or codes.

Table 6-4 provides recommendations to continue the Source modeling work of Chapter 5. Note that further investigation with both generic models and site-specific models will be part of the SARFIIP Salinity Management Measures.

Table 6-2 Recommendations for further work using the MODFLOW model of Chapter 4

Topic	**Recommendation**	**Reason**
Sensitivity analysis	Conduct sensitivity analyses to key parameters, including but not limited to: • Murray Group transmissivity, simulating the full saturated thickness of the Murray Group, rather than one aquifer unit • Solute transport parameters such as longitudinal and transverse dispersivity and porosity • Initial conditions for the salinity	Need to know whether these parameters are critical to the model results, or if they can be approximated with minimal data.
Other hydrogeological situations in the SA MDB floodplain	Simulate the circumstance where vertical upwards flow is the predominant source of regional groundwater flux into the floodplain aquifer, by increasing the potentiometric head in layer 4 and increasing the vertical hydraulic conductivity of the regional aquitard in layer 3.	Need to investigate whether the preliminary recommendations are applicable for regions such as Woolpunda and Waikerie.
Stress periods and ET	Trial adaptive stress periods for ET, similar to what has been investigated for the representation of changing river levels. Consider how to select stress periods which are based on both changing river levels and ET simultaneously.	Floodplain dynamics sensitive to ET.
River boundary condition	Trial the use of the stream package for the River Murray.	The stream package simulates a water balance for the river, so this may be a better representation. If successful, it would allow easier linking to surface water models such as Source.
Spatial discretization and floodplain geometry	Consider a scenario with a curving river.	The simulations so far indicate minimal sensitivity to spatial discretization. Need to determine whether this is an artifact of the highly simplified floodplain geometry.
Freshwater lenses	Use the density-dependent package SEAWAT to simulate freshwater lenses.	Current model assumes constant salinity with depth in the floodplain aquifer, whereas data shows the salinity changes with depth, especially where there are riverine freshwater lenses.

Table 6-3 Recommendations for groundwater modelling using other models and model platforms

Topic	**Recommendation**	**Reason**
Site-specific model	Apply the recommendations of Table 6-1 to a	The recommendations are derived from a model

	site-specific model, to see how they work in practice. (Note that this already is underway, DEWNR's development of the Pike Floodplain model.)	with a highly-simplified floodplain geometry and parameters. Are the recommendations still valid for complex site simulations?
Inundation recharge from natural floods and artificial/environmental watering	Trial the use of the 1D unsaturated zone simulation package for MODFLOW for the calculation of inundation recharge. Use results from a LEACHM model of the unsaturated zone (currently being developed for SARFIIP) to inform inundation recharge rates and zones.	Need to explore alternatives to representing inundation as specified recharge, to improve accuracy.
Net recharge (recharge and ET)	Trial the use of the MODFLOW ET package recently developed by Doble *et al.* (2015), which uses a lookup table for recharge with respect to land use, climate, soil, vegetation type, and depth to water.	The package is a compromise between fully-integrated models of the unsaturated and saturated zones. Unsaturated zone simulations are run in e.g. LEACHM or WAVES, and the results summarized in a lookup table that is read by MODFLOW during a simulation. May improve the accuracy of the model.
Local grid refinement	Trial codes which allow for local grid refinement, such as MODFLOW-USG and iMOD.	The codes allow a regional-scale model to be refined at a selected location. This would allow sub-models to be developed within a regional model, improving accuracy at a location while reducing model construction times and computational times.
Coupled modelling	Trial using GSFLOW, USGS's coupled surface water and groundwater flow model based on MODFLOW	A simpler alternative to fully-integrated models.
Fully-integrated modelling	Trial a fully-integrated model and compare results with a multi model approach, i.e. compare results with MODFLOW, Source and unsaturated zone models	Need to determine what circumstances warrant a fully-integrated modeling approach.
Rapid assessment tools	Evaluation of codes and methods that could be used to support a Salinity Risk Framework. Evaluation of codes and methods as rapid assessment tools for salinity impacts.	Need for a rapid tool to assess salinity impacts, as it is not always desirable or possible to develop a detailed MODFLOW model for this. Source is one possible tool.

Table 6-4 Recommendations for Source model

Topic	Recommendation	Reason
Wetland area-stage relationship	The code should be changed so that the same area-stage relationship is used for both leakage and evaporation calculations.	The current model routines assume a constant wetland area for calculation of leakage fluxes, but the evaporation routine used by Source assumes that wetland area varies with floodplain stage. This means that floodplains inevitably dry out by leakage rather than by evaporation, and hence wetland salinities are probably underestimated.
Groundwater velocity	Review and perhaps revise the methodology.	The salt transport routine incorporated into the Source model calculates groundwater flow using a constant velocity, and this results in salt loads to the river that are independent of the river level.
Comparison with MODFLOW	Run Source and MODFLOW with matching scenarios for a broad range of input parameters. Compare salt loads derived from the two models.	The most efficient means for evaluating some of the assumptions in the Source model, is by setting up and running parallel simulations using Source, and a more complex model such as MODFLOW. This would provide direct information on the accuracy of the approaches used in Source.

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