Spatial relationships between vegetation cover and irrigation-induced groundwater discharge on a semi-arid floodplain, Australia

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Received 16 June 2004; received in revised form 31 January 2006; accepted 3 February 2006

Summary Insight into spatial patterns of groundwater discharge through evapotranspiration is critical for understanding responses of groundwater dependent ecosystems. In the semi-arid floodplain environment of the lower River Murray, South Australia, patterns of steady state groundwater flux may indicate potential areas of anthropogenic salinisation due to adjacent irrigation and aid in management decisions for the protection of indigenous floodplain vegetation. This paper outlines the use of relationships for evapotranspiration response to groundwater depth and salt concentration, within the MODFLOW 2000 framework, to model spatial patterns of groundwater flux. This modelling approach is used to develop relationships between the steady state net rates evapotranspiration, seepage and baseflow, and floodplain properties including elevation, geometry, soil and vegetation distribution relevant to the field site.

The model was applied to a key research site on the River Murray, South Australia, to facilitate understanding of the factors contributing to the development of spatial patterns of irrigation induced groundwater discharge. The inclusion of elevation and vegetation data in the methodology revealed the effects of microtopography and plant water use on groundwater discharge. Spatial distribution of vegetation had the highest correlation with patterns of groundwater evapotranspiration. Floodplain elevation, floodplain width and geometry, which related to vegetation cover, also affected both groundwater discharge and baseflow to the river. © 2006 Elsevier B.V. All rights reserved.

Introduction

In arid and semi-arid regions, floodplains are ecologically and hydrologically important parts of a catchment. Due to their proximity to rivers, floodplains and their associated...
wetlands have higher biodiversity value than surrounding areas, providing habitat for many aquatic and riparian species, including vegetation, birds, mammals and fish species. In lowland gaining streams groundwater flowing from the catchment moves through the floodplain before either flowing to the river or being attenuated by evapotranspiration.\(^1\)

The shallow watertables found within floodplains mean that evapotranspiration and seepage in this environment is vital to produce an accurate water balance for the system. In semi-arid areas where saline regional groundwater discharges to streams, such as the lower River Murray in South Australia, evapotranspiration from rising floodplain watertables and altered flow regimes have led to the dieback of environmentally significant riparian vegetation such as red gum (*Eucalyptus camaldulensis*) and black box (*Eucalyptus largiflorens*) (Jolly et al., 1993). The naturally saline regional groundwater tables have risen as a result of higher river levels and a three orders of magnitude increase in recharge from adjacent irrigation developments, leading to the accumulation of salt in the soil and alluvial aquifer through evapoconcentration (Jolly, 1996). The hydrology, geomorphology and vegetation coverage of these floodplains are complex, leading to high spatial variability in evapotranspiration, and variation in vegetation health impacts.

Management of vegetation health is undertaken through pumping regional groundwater to lower riparian watertables, targeted spear point pumping under key eucalypt communities, injection of fresh water under communities, and localised artificial flooding or environmental irrigation. These management techniques require an understanding of spatial patterns of salinisation risk on a fine scale, and the true ecological impact of land management and irrigation decisions requires knowledge of the proximity of salinisation risk to vegetation communities. The ability to quantify spatial patterns of groundwater discharge through modelling and site characterisation is required in order to understand the potential impacts from salinisation on vegetation. To do this, a conceptual model of discharge that accounts for the variability in elevation, soils and vegetation cover on floodplains is needed.

Groundwater loss due to evapotranspiration, however, is known to be the most difficult component of the groundwater balance to estimate (Freeze and Cherry, 1979). Much soil science research of the 1950–1980s was focused on the understanding, quantification and representation of capillary upflow from shallow watertables (Gardner, 1958; Warrick, 1988), and more recently transpiration within the context of environmental science (Thorburn et al., 1995; Mac Nish et al., 2000) and silviculture (Silberstein et al., 1999; Feikema, 2000). Despite this understanding, numerical modelling of groundwater loss has tended to simplify the evapotranspiration process to the point of representing it as a simple sink with no interface definition, or as linear approximations.

Where patterns of groundwater evapotranspiration representing salinisation risk at a fine scale are of interest, however, simplified models of groundwater–atmosphere flux can limit the understanding of spatial variation and heterogeneities. A model that incorporates the key factors influencing evapotranspiration, including vegetation type, groundwater salinity, watertable depth and flooding frequency is required to investigate the origins of spatial patterns of groundwater discharge and salinisation.

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\(^1\) For the remainder of this paper, any reference to evapotranspiration refers to capillary upflow from shallow groundwater, rather than direct evaporation of rainfall.
Site description

Clarks Floodplain, adjacent to the Bookpurnong Irrigation District (Fig. 1), is located on the lower reaches of the River Murray in South Australia (34°21'S, 140°37'E). The field site is located within the semi-arid inland of Australia, with rainfall varying between 200 and 300 mm a⁻¹ and potential evaporation of approximately 1800 mm a⁻¹.

The lower River Murray floodplain is characterised by a flat, wide, meandering river within a deep river valley, excited during the early to middle, and then again in the late Pleistocene, and redeposited with coarse-grained quartz sands of the Monoman Formation during higher, post-glacial sea levels between 17,000 and 6000 BP (Twdale et al., 1978). In the late Holocene (approximately 4000 BP), the Coonambidgal Clay, typically consisting of clays and silts, was deposited in the river valley overlying and partly confining the Monoman Formation (Fig. 2). The modern floodplain surface tends to have rapid variation in soil type due to geomorphological processes of deposition and erosion that are continually occurring. Palaeochannels and backwaters formed from previous river courses are present in many parts of the floodplain.

The objectives of this study were therefore to model evapotranspiration at a scale and complexity appropriate for understanding salinity effects on vegetation health, and to use these results to understand the relative contribution of physical, ecological and hydrological parameters in governing groundwater flux distribution.

Specifically, the aims of the paper were to:

- Use relationships for evapotranspiration in semi-arid environments within a finite difference groundwater model in order to understand the long-term patterns of evapotranspiration and salinisation for the purpose of determining vegetation health in a floodplain environment; and
- Develop a conceptual understanding of the hydrological relationships between regional groundwater inflows, vegetation water use and baseflow to the river for a key research site to understand the factors that govern groundwater flow processes within a floodplain environment.

The first part of this study briefly describes the modelling process undertaken, including the use of a spatially distributed evapotranspiration relationship as an input for MODFLOW 2000 in order to facilitate better representation of groundwater discharge fluxes. This section provides a background to the modelling, and a more detailed description of this methodology is available in Doble et al. (2005).

The second section of this paper applies the model to a floodplain on the River Murray, South Australia in order to understand the floodplain hydrological processes. Model results are compared against spatial patterns of discharge from remote sensing images and fine scale baseflow data from geophysical measurements. A conceptual understanding of the relative importance of factors such as floodplain elevation, geometry, soil and vegetation type on the distribution of discharge is then developed for the field site.

Simulation of groundwater flux patterns

The intent of the modelling in this section of the paper was to use a better representation of evapotranspiration processes within a finite difference groundwater model to develop a conceptual understanding of the processes leading to spatial patterns of evapotranspiration. A more detailed discussion of the model development is given in Doble et al. (2005).

Conceptualisation

The methodology used in this paper involves representing spatial distributions of net groundwater flux by developing a relationship to use within a finite difference groundwater flow model, in this case MODFLOW 2000. This relationship uses spatial information for vegetation, soil and groundwater salinity distributions, as well as flooding duration and frequency to develop a spatially variable, depth-dependent ET function that can be used within MODFLOW. The function, which provides a surface boundary condition to the groundwater model, is similar to the segmented evapotranspiration function (ETS1, Banta, 2000) but also incorporates a recharge component, with the equilibrium depth, rather than
extinction depth describing a net flux of zero. The net rate of groundwater flux (recharge subtracted from evapotranspiration) is important for understanding the long-term or ultimate rate of salt accumulation, and is a common measure of salinisation potential (Petheram et al., 2003; Slavich et al., 1999). Similarly, the equilibrium depth, where the net flux is zero is critical for determining whether salinisation will occur. The model is run at steady state, in order to understand the ultimate salt accumulation potential for future riverine and vegetation management planning.

The intent of the study was not produce a model to accurately represent field conditions for management purposes, and to calibrate the model to produce a close fit with historical data. Instead, the aim was to conceptualise relationships between floodplain groundwater processes, leading to patterns of evapotranspiration. The modelling therefore required, rather than full calibration, only that the model parameters were within a reasonable range, likely to be found in the field situation being modelled. Spatial and numerical validation of the model was performed using a series of geophysical, remote sensing and field measurement techniques. A brief description of the spatial validation is given later in the paper in order to justify the model results.

**Previous solutions for groundwater discharge**

To determine a threshold for salinisation due to steady state evaporation of saline groundwater, Warrick (1988) proposed that the limiting rate of upward flow in a soil was inversely
proportional to the depth to the watertable but varied according to soil properties, as described by:

$$q_{\text{lim}} = \frac{A_n}{C_0 n}.$$ \hspace{1cm} (1)

From Eq. (1), the maximum moisture flux or diffuse discharge from a soil can be determined for varying groundwater depths. Shallower groundwater will lead to higher evaporation rates, and therefore greater concentration of salts in the surface soil.

The science of predicting both dryland and irrigation-induced soil salinisation using groundwater discharge rates has already been established in soil physics, but is less frequently described by groundwater hydrologists. Using relationships described by Gardner (1958), Wind (1955) and
others, Talsma (1963) and Peck (1978) defined thresholds for salinisation in terms of groundwater fluxes of $10^{-3}$ m d$^{-1}$ for irrigated areas and $10^{-4}$ m d$^{-1}$ for dryland conditions, respectively. This definition relates salinisation to rates of groundwater discharge and accounts for soil type and groundwater depth.

In numerical models, a linear function, or a constant flux is often used to represent evapotranspiration. MODFLOW 96 (McDonald and Harbaugh, 1996) assumes a linear evapotranspiration function where evapotranspiration varies from a maximum at the ground surface, to zero when the watertable is at some depth below the surface (extinction depth), such that:

$$q = q_{\text{max}}, \quad d \leq 0,$$

$$q = q_{\text{max}} \left(1 - \frac{d}{d_{\text{ext}}}\right), \quad 0 < d < d_{\text{ext}},$$

$$q = 0, \quad d \geq d_{\text{ext}}.$$

The linear nature of this relationship may cause discharge to be under-predicted where the watertable is deeper, and over-predicted where it is shallow, a problem when knowledge of groundwater discharge is required on a fine scale. MODFLOW 2000 includes the module ETS1, which represents evapotranspiration as a piecewise linear function (Banta, 2000), which is used to prevent this problem. Setting up an accurate curve, however, requires knowledge of the spatially varying floodplain conditions.

Eq. (1) describes unsaturated zone groundwater movement at a point scale. Difficulties arise in applying this model at a regional, or even local scale to consider spatial distribution of groundwater discharge. The linkage of point-scale unsaturated models with large-scale saturated groundwater flow models has been trialled extensively, but these models tend to have stability issues due to the sensitivity of the Richards Equation to lower boundary fluxes (Boone and Wetzel, 1996). Scaling up data from point scale unsaturated models to large-scale saturated groundwater models requires some means of extrapolation (Hatton, 1998) to match spatial and temporal scales, and the processes that are relevant to the system can change between small scales of the unsaturated model and large scales of the saturated model.

The methodology used in this paper employs an equation for groundwater discharge from plants developed for saline, semi-arid conditions, plus information on recharge rates and flood recurrence patterns, and applies it spatially to the MODFLOW 2000 ETS module using a geographic information system (GIS).

**Recharge and discharge**

An analytical model for the uptake of saline groundwater by plants in semi-arid regions was developed by Thorburn et al. (1995), based on unsaturated zone water and solute balances. In this model, discharge of groundwater by vegetation depends on soil water content, groundwater salinity and the salinity threshold at which vegetation can no longer extract groundwater derived soil water. This relationship for instantaneous flux is described by the equation:

$$U = A \left(\left[Z_g[1 - \gamma]\right]^{n+1} + \frac{(n + 1)A_C t_d}{\theta_i C_{th}}\right)^{\frac{1}{n+1}} \left(1 - \frac{C_g(\theta_i - \theta_f)}{\theta_i C_{th}}\right).$$

(3)

The cumulative groundwater discharge ($D$) through plant water uptake is found by integrating Eq. (3) over the time period from $0$ to $t$:

$$D = \left(\left[Z_g[1 - \gamma]\right]^{n+1} + \frac{(n + 1)A_C t_d}{\theta_i C_{th}}\right)^{\frac{1}{n+1}} \left(1 - \frac{Z_g[1 - \gamma]}{Z_c}\right)$$

$$\times \left(\theta_i \left(\frac{C_{th}}{C_g} - 1\right) + \theta_f\right).$$

(4)

For a floodplain environment, this represents the cumulative discharge occurring for the continuous period between flood cycles, where $t_d$ represents the length of the dry period between the end of one period of inundation ($t = 0$) to the start of the next ($t = t_f$). The length of this dry period is critical for vegetation dependent on groundwater, as evapotranspiration processes will concentrate salt, which in turn limits plant water uptake. When an overbank flow event occurs, recharge from flooding will freshen soil and ground water, allowing plant water uptake to increase. More frequent flooding (shorter $t_d$) will therefore support higher volumes of discharge in the long term than would be possible with longer dry periods.

Recharge into the floodplain will be dependent on both the vertical conductivity of the soil and the period for which that part of the floodplain is inundated. This volume of infiltration, less the water volume that fills the unsaturated zone to field capacity, gives the net volume of water recharging the groundwater. Recharge may be represented by:

$$R = K_{\text{sat}} t_f - ((\theta_{sc} - \theta_f) Z_f + (\theta_{sc} - \theta_i) (Z_g - Z_f))$$

$$\text{if } (\theta_{sat} - \theta_{sc}) Z_g > 0.$$

(5)

For simplification, the moisture content within the capillary fringe ($\theta_i$) may be approximately equal to the moisture content at field capacity ($\theta_{sc}$), therefore the second term will cancel. Groundwater recharge in conductive soils will also be limited by the total unsaturated pore space that may be filled during floods. The equation for recharge is therefore the minimum of the following equations:

Minimun: $R = K_{\text{sat}} t_f - (\theta_i - \theta_f) Z_f,$

$$R = (\theta_{sat} - \theta_i) Z_g,$$

where $(\theta_{sat} - \theta_i)$ represents the pore space available to receive recharge water.

Combining recharge and discharge for a cycle of one flood and one discharge period, an approximation of the net groundwater flux from plant water use ($q$) is:

$$q = \min \left\{ \frac{K_{\text{sat}} (t_f) - (\theta_i - \theta_f) Z_f}{(\theta_{sat} - \theta_i) Z_g} - \left(\left[Z_g[1 - \gamma]\right]^{n+1} + \frac{(n + 1)A_C t_d}{\theta_i C_{th}}\right)^{\frac{1}{n+1}} - Z_g[1 - \gamma]\right\}$$

$$\frac{(\theta_i \left(\frac{C_{th}}{C_g} - 1\right) + \theta_f)}{t_f + t_d}. $$

(7)
Eq. (7) reflects the long term balance between groundwater discharge during dry periods, and recharge during floods. The relationship contains information on evapotranspiration from groundwater dependent ecosystems, and flood frequency and duration patterns, evaluated as a long term trend to determine whether areas are dominated by recharge or discharge, dependent on depth to groundwater. In areas with saline regional groundwater, discharge dominance will lead to secondary salinisation.

This equation describes a relationship between groundwater flux and watertable depth that can be used within a finite difference groundwater flow model as a surface boundary condition. The function was used to calculate input parameters for the ETS1 module using GIS coverages of elevation, soil and vegetation conditions for each part of the floodplain and each model cell. Data used for this spatial calculation is shown in Table 1, and durations of inundation and dry periods ($t_d$ and $t_r$) provided from flood recurrence data.

**Seepage**

At many points along the river valley, the high incoming regional groundwater inflows cause the watertable to intersect the ground surface at the break of slope, causing direct seepage of groundwater. Seepage at the break of slope was modelled with an additional segment of the evapotranspiration function, such that the maximum ET rate at the soil surface was equal to the potential evapotranspiration for the region (Table 2). At a point 0.2 m below the surface, the evapotranspiration function reverted to the groundwater flux relationship appropriate for the soil and vegetation conditions.

**Confining layers**

For Clarks Floodplain, and most areas of the lower River Murray, the Coonambidgal Clay forms a confining layer over the Monoman Sands aquifer. The effect of this confining layer is to increase the pressure head required to force groundwater to the surface, or high enough that the capillary fringe reaches the surface for evaporation to occur. This process may be modelled within MODFLOW using a false evaporation surface above the actual ground surface to represent this additional head requirement. This stretches the recharge–discharge curve both upwards above the equilibrium point, and downward within the recharge section below the equilibrium point.

The degree of upward stretching is given by the Darcy equation, which specifies the head required for maximum groundwater flux at the surface. That is:

$$ q = K \frac{\Delta h}{L}. $$

Rearranging and substituting into Eq. (4) for the condition of the watertable at the soil surface, and therefore $K = K_{sat}$ gives:

$$ \Delta h = \frac{L}{K_{sat}} ET_{max}. $$

where recharge is zero, or:

$$ \Delta h = \frac{L}{K_{sat}} \left( \left( \frac{A}{(n+1)AC_{th}a} \right)^{1/\alpha} \left( \frac{C_{th}}{C_{sat}} - 1 \right) \right). $$

For the groundwater flux curves, the maximum rates of evapotranspiration and recharge remain the same.

**Coupling with MODFLOW**

The relationships described in the previous sections are applied to MODFLOW 2000 (Harbaugh et al., 2000) through a modified version of the ETS module (Banta, 2000), which combines both recharge (RCH) and discharge (ETS) together as a single groundwater flux dependent on groundwater depth (Fig. 4). The resulting 'combined recharge–discharge' (CRD) module is a simple additive combination of these two functions.

The function applies a net recharge or discharge to the model, depending on whether the groundwater head is above the ground (max ET), between the surface and...
equilibrium depth (a) (net ET), between the equilibrium depth and extinction depth (b) (net recharge) or below the extinction depth (max recharge). The point at which groundwater discharge is equal to zero is termed the equilibrium depth, that is where recharge and discharge fluxes are equal in the long term. With this method of defining groundwater flux it is possible to include varying recharge functions such as zero recharge when groundwater is at the surface.

Vegetation input parameters were applied spatially using a GIS vegetation survey (Telfer and Overton, 1999), and soil and groundwater parameters using a map of soil types derived from soil logs and geomorphology, and groundwater chloride concentration data. Each of the three coverages was divided into categories shown in Table 1 and assigned values for parameters used in Eq. (7). A GIS framework was used to calculate four points on the CRD curve, defining maximum ET, maximum recharge, and equilibrium depth. Spatial distributions of maximum ET, maximum recharge and equilibrium depth were used as inputs for MODFLOW, which provided the hydrological component and groundwater depth required to calculate long term groundwater flux.

### MODFLOW model development

Model parameters and their sources are listed in Table 2. Fig. 5 shows an example cross-section of the model, including the geological layers used. On the highland, the upper, unconfined aquifer consisted of Upper Loxton Sands ($K_h = 5 \text{ m d}^{-1}$), which was underlain by the Lower Loxton Sands ($K_h = 0.2$). On the floodplain, the alluvial aquifer within the Monoman Formation ($K_h = 10$) is again underlain by Lower Loxton Sands. Overlying the alluvial aquifer is a spatially variable semi-confining layer, consisting of heavy Coonambidgal Clay (K variable) at the break of slope, to a silty sand deposit near the river. Thicker Coonambidgal Clay was found in some low lying palaeochannels. Vertical hydraulic conductivity and other soil properties used for the evapotranspiration function (Eq. (7)) are shown in Table 1.

The MODFLOW model was discretised into a regular, square grid with cell dimensions ranging from 200 × 50 m

<table>
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<th>Table 2 Model parameters</th>
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<tr>
<td>Parameter</td>
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<tr>
<td>Recharge and discharge parameters</td>
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<tr>
<td>Irrigation recharge</td>
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<td>Dryland recharge</td>
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<td>Maximum evapotranspiration rate on floodplains</td>
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<td>Open water evaporation rate</td>
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<td>Seepage extinction depth</td>
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<td>River and elevation parameters</td>
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<td>River stage upstream of Lock 4</td>
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<td>Floodplain elevation</td>
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<td>Upper Loxton Sands</td>
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to $6.25 \times 6.25$ m, refined about the break of slope, as indicated in Fig. 1. The model used constant head cells on the outer boundaries with elevations of 12.5 m, 13.9 m, 15.2 m and 12.5 m aHD, starting from the north-western corner and progressing clockwise around the corners of the model domain. Floodplain elevation data was available in spatial form at vertical intervals of 100 mm on a $30 \times 30$ m grid cell scale (Fig. 6, Overton, 2005). The river boundary was simulated using the MODFLOW river module. River stage was set at an elevation of 10.5 m below the weir and 13.2 m above it. The river is regulated into a series or weir pools, and river stage variation is minimal for the length of the floodplain. The model was run in steady state for current irrigation conditions, both for confined and unconfined alluvial aquifer states.

Vegetation and soil parameters from Table 1 were adopted from Thorburn, 1996, and extrapolated to sandier soils using Gardner $n$ parameters provided by Jolly et al. (1993) from field experiments. Groundwater parameters were taken from field measurements in March 2003. The spatial distribution of soil, vegetation and groundwater parameters are shown in Fig. 7a–c.

Historical flood data was used to determine the average return period between floods, and the duration of inundation for various river stages. These values were then used in Eq. (7). Data for River Murray flows into South Australia was obtained for the years 1957–2003 (post regulation) from gauging station GS426200SA (Porter, 2002). A rating curve measured at a point downstream of Lock 4 corresponding to Clarks Floodplain was used to link flow...
Figure 7  Distribution of: (a) groundwater chloride concentrations (g L\(^{-1}\)) for Clarks Floodplain obtained from field data, (b) soil type distribution, and (c) surveyed vegetation type, used as spatial inputs for the combined recharge–discharge function. Parameter values associated with these distributions are shown in Table 1.
frequency data to the spatial coverage of floodplain elevation (Fig. 6). Fig. 8 shows the resulting smoothed relationship for average period of inundation and duration between flooding as a function of floodplain elevation. This means of including flood information allows the long term trends of the recharge–discharge balance to be investigated.

Spatial distributions of the MODFLOW CRD module inputs: maximum ET, equilibrium depth and maximum recharge, were produced using Eq. (7), and the spatial data described above. Fig. 9 shows examples of various recharge–discharge relationships that cover several different vegetation, soil and groundwater scenarios on floodplain. The locations at which data for these curves were selected are indicated in Fig. 7a.

The MODFLOW model was run twice more using uniform evapotranspiration and elevation parameters, the results of which were used to determine how much additional information was gained by considering variation of the floodplain system at a finer scale. The first run consisted of a constant floodplain elevation of 14 m (herein referred to as the uniform model), and the second with a spatially variable surface elevation (variable elevation model, Fig. 6), but with evapotranspiration not defined by vegetation parameters. Both models used a constant maximum ET rate of 0.3 mm d\(^{-1}\) and an equilibrium depth of 3.5 m, zero floodplain recharge and a constant vertical hydraulic conductivity (0.01 m d\(^{-1}\)).

**MODFLOW model results**

A spatial distribution of net groundwater flux was produced using the methodology described above (Fig. 10). This was compared with the flux outputs from the uniform (Fig. 11) and variable elevation (Fig. 12) models. Inclusion of elevation, vegetation, soil type and flooding information within the MODFLOW model provided finer-scale predictions for discharge patterns and revealed a variation in discharge rates across the floodplain. Not only could areas at risk of salinisation be identified, but variations in the rate of salinisation were also quantifiable.

The MODFLOW groundwater flow model predicted high net discharge of groundwater at the back of the floodplain, near the break of hill slope where the depth to groundwater was at a minimum (Fig. 10). High groundwater discharge was also associated with red gum and black box forests at the ends of floodplain peninsulas and adjacent the river in divisions 1, 7 and 8. Flow divisions used to describe discharge locations are numbered in Fig. 1. Localised elevation, or microtopography, was revealed by high evapotranspiration rates in the spatial patterns of groundwater discharge, both at the break of slope, and associated with the low-lying palaeochannel that extended along the centreline of Clarks floodplain in division 7. Patterns of elevation were evident within vegetation classes, for example comparing the elevation profile at the river end of division 3 (Fig. 6) with the groundwater discharge patterns in Fig. 10. The effect of elevation between vegetation communities was not evident.
Validation
In light of the objectives of this paper, the validation process described below provides a means of establishing that both the modelled spatial patterns of evapotranspiration and the rates of groundwater flux lie within a range that is appropriate for the field conditions under investigation. Since the model was not used for detailed floodplain management, a thorough calibration with historical data was neither necessary nor productive. This validation is therefore intentionally concise and indicative in order to give confidence that the processes representing evapotranspiration are adequate, rather than providing detailed calibration results.

Higher rates of salinisation from evapotranspiration at the break of slope and through the centre of the floodplain are indicated by dead and unhealthy vegetation in Fig. 3, as well as observations of seepage and significant accumulation of salt in the soil profile during field investigations.

Point scale transpiration measurements from the field were compared with modelled results. Sap flow measurements from Thorburn and Walker (1994), Jolly et al. (1991) and Akeroyd et al. (1998) indicate transpiration rates within the lower River Murray region from 70 to 700 mm a⁻¹, and 35 to 150 mm a⁻¹ for red gum and black box eucalypt species, respectively. Evapotranspiration from bare soils on Chowilla Floodplain in the same region of the River Murray was approximated using ²H isotope profiles to be 10 mm a⁻¹ (Jolly et al., 1993) with a groundwater depth of 4–5 m. Accounting for recharge, modelled ET rates were approximately 180–360 mm a⁻¹ for red gum areas,
40–130 mm a\(^{-1}\) for black box and 0–40 mm a\(^{-1}\) for the elevated grassland, lying within the ranges described. A shallower watertable (2–3.5 m) may explain why the grassland evaporation rates were slightly higher than that described by Jolly et al. (1993).

Satellite imagery was used as an indicator of vegetation water use (Goodrich et al., 2000). Spectral indices such as normalised difference vegetation index (NDVI) that are derived from satellite images can indicate vegetation distribution and health or density changes due to salt accumulation. In groundwater dependent ecosystems, they are also able to provide a direct indication of groundwater discharge through transpiration, which may be correlated with groundwater discharge model outputs. NDVI and other vegetation water use indices indicate plant water use only, however, and do not account for direct soil evaporation.

Landsat-5 TM images with a 25-m resolution were obtained for the 18th February 2002, and the NDVI was calculated using the normalised difference of the red and infra red spectral bands (bands 3 and 4). This data was not used to create or calibrate the model. Fig. 13 shows the spatial distribution of NDVI across the floodplain, with darker areas indicating high water use. The image indicates high water use at the ends of the floodplain peninsulas and adjacent...
the river. High water use is also evident along the break of slope, indicating areas where fresh irrigation recharge has broken through to the floodplain seepage face and is providing a permanent water supply to waterlogging tolerant vegetation such as grasses and reeds. There were strong correlations in high water use at the ends of the peninsulas and adjacent the river in both model output and NDVI. This indicates that the modelled discharge function is providing an adequate basis for further analysis of spatial patterns of discharge.

NDVI values and long term discharge fluxes from the model output were compared within different eucalypt vegetation categories on Clarks Floodplain (Fig. 14, Table 3). Areas where evaporation from annual species or bare soil dominated, such as the break of slope and floodplain centre, were excluded from this analysis. Within the tree categories, a trend of increasing modelled discharge rates with higher NDVI values was observed, suggesting a positive relationship between modelled and indicated discharge.

To test the hydraulics of the model, baseflow to the river was tested against field measured and indicated baseflow.

![Comparison of Modelled Discharge and NDVI](image)

**Figure 14** Comparison between modelled long term ground-water discharge and NDVI for Clarks Floodplain for areas with eucalypt vegetation. Vegetation classifications are found in Table 3.

Monitoring of river salinity has estimated that Clarks Floodplain contributes salt loads of between 30 and 50 t d⁻¹ (Porter, 2002). This also correlated with modelled salt loads from Clarks Floodplain, which were equivalent to 45 t d⁻¹.

Transient electromagnetics at a nanometre scale (Nano-TEM) has been used to measure resistivity in river sediments along the stretch of river between Lock 3 and 4 of the River Murray in South Australia (Berens et al., 2004) in order to identify the conductive anomalies associated with concentrated saline groundwater inflows. NanoTEM analysis does not provide actual values of salt loads or baseflow, but provides a qualitative means of comparing the relative ground-water inputs to a watercourse.

Modelled baseflow showed higher groundwater discharge in areas where the river is close to the highland, and negative baseflow at the ends of the peninsulas, correlating strongly with the nanoTEM resistivity patterns. Multiplying baseflow rates by a regional groundwater salinity of 30,000 mg L⁻¹ resulted in values of modelled salt inputs.

Modelled salt loads at approximately 250 m intervals were compared quantitatively with nanoTEM results using a threshold resistivity of 27 Ω m to distinguish positive and negative inflows (Fig. 15). Patterns of high conductivity correlated well with the modelled baseflow.

A map of modelled depth to groundwater is provided in Fig. 16, compared with groundwater depth measured from alluvial groundwater wells. Root mean square error of modelled vs field groundwater elevation was calculated to be 0.59 m, with the data biased toward the lagoon and break of slope where the majority of boreholes were located. The greatest variation between field and modelled heads was found between the lagoon and break of slope where the change in hydraulic gradient was greatest. Whilst the depth to groundwater provides an opportunity for a standard model calibration and indicates the underlying groundwater conditions, it does not provide information on the distribution of vertical groundwater flux, as there is no provision for spatially variable soil, vegetation or groundwater salinity conditions. Measuring the potential for salinity as depth to watertable will be of little benefit if the spatial distribution of salinisation potential is of interest.

![Salt Loads: nanoTEM and Modelled](image)

**Figure 15** Salt load patterns from Clarks Floodplain showing modelled data and resistivity from the nanoTEM information using a threshold value of 27 Ω m to distinguish positive and negative baseflow, and linearly scaling conductivity to salt loads.

<table>
<thead>
<tr>
<th>Classification</th>
<th>Vegetation type</th>
</tr>
</thead>
<tbody>
<tr>
<td>1r(f)</td>
<td>Healthy red gum forest</td>
</tr>
<tr>
<td>1br</td>
<td>Healthy mixed black box red gum</td>
</tr>
<tr>
<td>1b</td>
<td>Healthy black box woodland</td>
</tr>
<tr>
<td>1bl</td>
<td>Black box with lignum understory</td>
</tr>
<tr>
<td>1r(r)</td>
<td>Red gum regeneration</td>
</tr>
<tr>
<td>2r</td>
<td>Unhealthy red gum woodland</td>
</tr>
<tr>
<td>2r1bc</td>
<td>Unhealthy red gum, healthy black box and cooba</td>
</tr>
</tbody>
</table>

---

**Table 3** Classifications for the vegetation survey of Clarks Floodplain, used in the NDVI analysis (Fig. 14) for model validation.
Spatial patterns of high groundwater discharge

While Fig. 10 shows a distribution of groundwater flux similar to that found from vegetation indices, the simpler uniform and variable elevation MODFLOW models indicated high discharge associated with the break of slope (Figs. 11 and 12), but none of the groundwater flux associated with vegetation water use. Some discharge was predicted on the low-lying edges of the floodplain for the variable elevation model run, however, and higher evapotranspiration rates are evident around the palaeochannel in the centre of the floodplain. While these simplifications may be sufficient for broad regional scale modelling, the lack of distributed vegetation data meant that representation of groundwater discharge was less accurate on a sub-floodplain scale.

Table 4 shows a budget of groundwater flow for each floodplain division. As a general trend, groundwater inflow from the highland increased for each division from the northwest to the southeast depending on the proximity of the break of slope to the highest point of the groundwater mound. Seepage only occurred within divisions 5–10, while evapotranspiration was present on all divisions, and was highest for divisions 1, 7 and 8. Baseflow was negative for four of the divisions (1, 2, 7 and 8), indicating flow from the river into the alluvial aquifer.

The areas in which the net groundwater flux is recharging the aquifer are predicted by the model to be minimal, and occur only in very low lying areas on the ends of floodplain peninsulas where sandy deposits occur and vegetation is yet to establish. The dominance of ET within the system is expected to be due to the significantly decreased flood recurrence interval from river regulation in the lower River Murray. Net groundwater recharge from flooding is predicted to occur predominantly in floodplain divisions 1, 3, 7 and 8, characteristically the low-lying and sandy inside of meanders (Table 4).

Table 5 shows the width, amplification factor, average elevation, average hydraulic conductivity and area of trees for each floodplain division. The area of trees indicates the area of floodplain supporting black box and red gum eucalypts, which have high water requirements compared with bare ground, lignum or grassland vegetation. The
description of area of trees in the following sections define low tree area as 0–10 ha of trees, medium tree area as 10–30 ha and high tree area as more than 30 ha of trees.

The spatial patterns of groundwater discharge tended to be quite robust, and did not change significantly when the groundwater discharge relationship was altered. The quantitative rate of groundwater discharge was far more sensitive to changes in the discharge function than the spatial patterns themselves.

Confined and unconfined aquifer responses

The evapotranspiration output for which surface confining effects were not modelled (Fig. 18) retains the overall spatial effects of the confined case (Fig. 10). Groundwater discharge is still predominant at the break of slope, ends of peninsulas and in the low-lying palaeochannel, but is higher than for the confined case. Seepage is especially more pronounced at the break of slope, as this area tends to be represented by a heavier clay with lower vertical conductivity, and is therefore more confining of vertical groundwater movement. The confining effects of the Coonambidgal Clay had little impact on the spatial patterns of groundwater discharge other than reducing the rates of groundwater seepage with respect to overall discharge rates. Groundwater discharge was lowered overall, but patterns associated with vegetation types and microtopography were retained.

Table 5

<table>
<thead>
<tr>
<th>Division</th>
<th>Floodplain width (m)</th>
<th>Amplification factor (–)</th>
<th>Area trees (ha)</th>
<th>Mean height above river at break of slope (m)</th>
<th>$K_v$ average (m d$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1492</td>
<td>12.36</td>
<td>39.22</td>
<td>3.91</td>
<td>0.0085</td>
</tr>
<tr>
<td>2</td>
<td>496</td>
<td>2.00</td>
<td>14.64</td>
<td>4.13</td>
<td>0.0034</td>
</tr>
<tr>
<td>3</td>
<td>830</td>
<td>3.90</td>
<td>16.44</td>
<td>3.99</td>
<td>0.0042</td>
</tr>
<tr>
<td>4</td>
<td>461</td>
<td>2.00</td>
<td>4.78</td>
<td>3.96</td>
<td>0.0048</td>
</tr>
<tr>
<td>5</td>
<td>494</td>
<td>1.11</td>
<td>3.60</td>
<td>3.42</td>
<td>0.0033</td>
</tr>
<tr>
<td>6</td>
<td>626</td>
<td>0.79</td>
<td>8.44</td>
<td>3.13</td>
<td>0.0040</td>
</tr>
<tr>
<td>7</td>
<td>2059</td>
<td>11.17</td>
<td>52.44</td>
<td>2.90</td>
<td>0.0045</td>
</tr>
<tr>
<td>8</td>
<td>1895</td>
<td>9.33</td>
<td>36.68</td>
<td>3.41</td>
<td>0.0028</td>
</tr>
<tr>
<td>9</td>
<td>544</td>
<td>1.27</td>
<td>2.04</td>
<td>3.35</td>
<td>0.0018</td>
</tr>
<tr>
<td>10</td>
<td>261</td>
<td>0.93</td>
<td>2.59</td>
<td>3.56</td>
<td>0.0028</td>
</tr>
<tr>
<td>11</td>
<td>119</td>
<td>1.00</td>
<td>2.54</td>
<td>3.70</td>
<td>0.0017</td>
</tr>
</tbody>
</table>

The distribution of groundwater discharge with distance from the break of slope (Fig. 17) confirms that while the unconfined scenario gives higher rates of groundwater discharge across the entire floodplain, the difference is more pronounced in the first 200 m from the break of slope where seepage occurs. In addition to the smaller vertical conductivity in this location, the confining effects from the Coonambidgal Clay are most significant where the watertable is close to or above the ground surface, such as at the break of slope, as the distance of upward fluid flow through clay before evaporation occurs is greater.

Modelling the surface confining layers may not be necessary in situations with deeper groundwater or coarser soils, but should be considered in situations where seepage or high discharge rates from very shallow groundwater are of interest.

Interpretations of model results

The remaining sections of this paper are directed toward the development of relationships between evapotranspiration and factors such as floodplain geometry, elevation and distribution of vegetation, soil parameters.

Surface elevation is a key parameter in determining where salt accumulation will occur, as the greatest rates of ground-
water discharge by evapotranspiration will be located in low lying areas where the depth to watertable is a minimum. Microtopography, or small-scale variation in elevation, defines both the areas of groundwater discharge and the frequency of flood inundation. As ecosystem communities are linked with microtopography, it also defines the impact on biodiversity in the region (Cramer and Hobbs, 2002).

Soil type (Talsma, 1963; Warrick, 1988), proximity to the river and vegetation characteristics (Holland, 2002) will also affect the extent of salt accumulation. The episodic leaching of salt from the soil profile by flood-induced recharge further complicates the establishment of salt stores. Floodplain geometry will affect distribution of groundwater flow to a river or water body, as observed in the embayments of Green Bay and Lake Michigan (Cherkauer and McKerghan, 1991), and meanders of the Swan River, Perth (Linderfelt and Turner, 2001). The sinuosity of the river, and associated geometry of the floodplain is also likely to determine the distribution of groundwater discharge from a floodplain environment.

It was hypothesised that each of these parameters would affect the distribution of long term groundwater flux and therefore salt accumulation. The order of importance of these parameters for defining discharge patterns, however, was unknown.

**Analysis**

The floodplain was divided into a series of flow divisions (Fig. 1) in order to compare characteristics of the floodplain within each division. The flow lines were defined as the flow path that groundwater took from a set of equally spaced points along the edge of the floodplain to the river, and were determined by generating a set of equipotential contours with groundwater flow being perpendicular to these and toward the river. This minimised groundwater flow between flow divisions. Zonebudget (Harbaugh, 1990) was used to perform water balance calculations for each of these divisions.

The average rate of evapotranspiration for a division, calculated at steady state, was compared with average elevation above the river, floodplain width and vertical hydraulic conductivity. Amplification or divergence of the division, that is, the length of river boundary divided by the width of the division at the break of slope was also compared with evapotranspiration and baseflow, as was the proportion of healthy and total trees found in each division. These results were compared with those from the uniform model. Results from the elevation only example were almost identical to the uniform model, and are therefore not shown.

**Dependence of seepage on local elevation**

Constriction of groundwater flow as it moves from the unconfined highland into the partially confined floodplain aquifer led to groundwater heads at the break of slope exceeding the floodplain elevation and caused seepage to occur. Seepage was greatest in the area of the palaeochannel, which corresponded to the points of lowest elevation along the cliff. These predictions match with observed seepage at the site.

There was very little correlation between the fraction of inflow from the highland expressed as seepage and the mean floodplain elevation for each of the flow divisions ($R^2 = 0.12$). Comparison with the mean elevation of a 200 m wide strip of floodplain at the break of slope where groundwater heads are highest however, gave an inverse correlation ($R^2 = 0.61$; Fig. 19). No grouping of divisions with high tree areas or wide floodplain (points 1, 7 and 8) were found, suggesting that, as expected, vegetation type had far less effect on seepage than elevation or geomorphology. Seepage is an elevation controlled process, and is most affected by low depths to watertable, whether by low surface elevation, or high groundwater table.

While the volume of seepage is dependent on inflow rates, seepage only appears in divisions with break of slope elevations lower than 3.7 m above the river stage. Apart from changes in hydraulic head, and therefore depth to

Figure 18  Predicted groundwater discharge on Clarks Floodplain, unconfined model.
groundwater, no other geometric factor significantly affected the seepage rate from the floodplain.

### Effects of vegetation coverage and floodplain width

The spatial distribution of vegetation across the floodplain had the greatest overarching effect on patterns of groundwater discharge. The area of high water using trees within each floodplain division correlated very closely with evapotranspiration ($R^2 = 0.96$; Fig. 20). While the maximum rate of groundwater discharge was found at the break of slope, associated with shallow groundwater and seepage, the total volume of discharge from the eucalypt communities was also high due to the large areas of vegetation with high water requirements and deep extinction depths.

Wider sections of floodplain result in a longer groundwater travel time within the floodplain. As the hydraulic head is constrained at both the break of slope and the river, longer flow paths reduce the hydraulic gradient and groundwater velocity. This consequently subjects groundwater to evapotranspiration for longer periods of time. Larger floodplain widths also provide a greater area over which evapotranspiration may take place. Fig. 21 shows a strong relationship between the rate of evapotranspiration from a division and the floodplain width ($R^2 = 0.96$) despite varying groundwater inflows from the highland. However, the area of trees within a division was also found to increase with floodplain width, indicating some interdependence.

Results from the uniform MODFLOW model showed no relationship with either tree area, which was expected due to the lack of vegetation parameters, nor floodplain width, despite the inclusion of floodplain spatial geometry within the uniform model. The lack of relationship between floodplain width and evapotranspiration in the uniform model confirms that ET is more dependent on vegetation than floodplain width alone.

### Dependence of groundwater discharge on floodplain elevation and soil type

Warrick (1988), Gardner and Fireman (1958) and Thorburn et al. (1992) show that the relationship between watertable depth and evaporation in soils is an inverse power relationship. It was hypothesised that given a relatively level piezometric surface, floodplain elevation, which varies more rapidly in a spatial context than groundwater elevation, is likely to be the primary variable for depth to watertable. Relative elevation of the floodplain is therefore thought to govern distribution of groundwater discharge on a spatial scale.

Fig. 22 shows evapotranspiration rates plotted against floodplain height relative to the river level. While there are no strong overall trends, there are sub-groups within the data set defined by the area of trees present within the division. The divisions with a high tree area exclusively represent the group with the highest ET rates. The second group, below this, is represented by divisions with a low to medium area of trees, and shows a trend of increasing ET with decreasing elevation ($R^2 = 0.64$).

The location of low-lying regions within a division will also dictate the relative effect of surface elevation. Low-lying areas that coincide with the break of slope (divisions 5–9) or eucalypt communities (ends of divisions 1, 3, 7
Spatial relationships between vegetation cover and irrigation-induced groundwater discharge

and 8) are likely to produce higher discharge rates than if they coincide with grassland or low water using lignum (*Muehlenbeckia cunninghamii*) covered areas further from the break of slope (centre of division 8). Within each vegetation class, elevation is also a factoring determining patterns of groundwater discharge. The effects of localised elevation patterns (Fig. 6) are visible within different vegetation areas, rather than between them, but as elevation is an indicator of flooding frequency, vegetation distribution and floodplain elevation are not completely independent variables.

The weaker dependence on elevation than expected is potentially as a result of the higher recharge rates associated with lower lying areas. Although lower elevation leads to higher rates of evapotranspiration, it is partly offset by increased recharge through more frequent inundation, so the sensitivity of net flux to elevation is lessened.

Groundwater discharge and average vertical hydraulic conductivity ($K_v$) of a division appear to have a limited relationship for Clarks Floodplain ($R^2 = 0.12$ for increasing discharge with increasing $K_v$) as shown in Fig. 23. Separating two groups of data points, high and low tree area, reveals a slight positive trend in discharge with increasing $K_v$, but this overall trend is overwhelmed by the effects of vegetation type on ET.

**Effects of floodplain amplification on baseflow**

The ratio between river frontage of a floodplain division to its width at the break of slope, or amplification factor, strongly influenced where baseflow to the river occurred. High amplification factors were associated with negative baseflow rates (Fig. 24), and within the divisions with low areas of trees, baseflow increased with both higher groundwater inflows and lower amplification rates. Divisions with long river frontages coincided with a wider floodplain and greater area of trees, facilitating high evapotranspiration rates adjacent the river. To achieve this evaporation rate, river water infiltrates into the alluvial aquifer, facilitated by the lower groundwater heads under the eucalypt communities. This behaviour has been recorded in riparian environments in the San Pedro River, described by MacNish et al. (2000) and Stromberg et al. (1996). While there was a notable correlation between baseflow and division amplification or river frontage, the area of trees present on a floodplain division was also represented in data clusters. The interdependent nature of river frontage and vegetation coverage made these parameters difficult to distinguish, and suggested that baseflow was defined by a combination of river geomorphology and ecology, rather than any single parameter.

Fig. 24 shows that the uniform model results were similar to the more detailed modelling for floodplain divisions with low amplification factors and low tree cover. Baseflow followed the same trend, increasing with rising highland inflows. For divisions with high amplification factors, however, the lack of spatial variation in the ET function meant the uniform model still predicted positive baseflow. Baseflow was shown to be influenced firstly by groundwater inflows to the floodplain, then by amplification factor and therefore also by proximity to the highland. Spatially variable ET rates were only important for areas further from...
the highland. This indicates that while patterns of ET are sensitive to spatially variable model inputs, baseflow patterns are relatively robust.

Regression

A linear multiple regression was undertaken, comparing the modelled evapotranspiration with highland inflow, floodplain width, area of trees, amplification factor, average $K_v$, and mean height of floodplain section above the river level. This indicated the relative contributing factors to evapotranspiration, and allowed a series of variables to be compared with evapotranspiration simultaneously. A relationship for evapotranspiration was derived with an $R^2$ value of 0.998. In order of significance, the parameters contributing to evapotranspiration were the area of trees, highland inflow, floodplain width and mean height above the floodplain, with $P < 0.05$ indicating significance (Table 6). Amplification factor and average $K_v$ had little effect on patterns of evapotranspiration. Baseflow, floodplain width and tree area were highly interdependent, with wider floodplain divisions also having high tree coverage and negative baseflow. Conversely, narrow divisions tended to have low tree coverage and higher, positive baseflow.

Results from the regression supported the conclusions drawn from the figures above.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>$P$-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Highland inflow (m$^3$ d$^{-1}$)</td>
<td>0.015</td>
</tr>
<tr>
<td>Floodplain width (m)</td>
<td>0.020</td>
</tr>
<tr>
<td>Area trees (ha)</td>
<td>0.003</td>
</tr>
<tr>
<td>Amplification factor (–)</td>
<td>0.325</td>
</tr>
<tr>
<td>Mean height above river (m)</td>
<td>0.041</td>
</tr>
<tr>
<td>$K_v$ average (m d$^{-1}$)</td>
<td>0.904</td>
</tr>
</tbody>
</table>

Table 6  Regression results for evapotranspiration

Net flux output from the model can also be used to understand the spatial patterns of salt accumulation. Salt accumulation rates were predicted by multiplying the coverage of groundwater discharge rates (Fig. 10) with a spatial map of local groundwater salinity determined from field studies. High rates of salt accumulation are predicted at the break of slope, and also on the outsides of the bend in divisions 3–7 (Fig. 25). Although Fig. 10 shows a high discharge rate from red gum and black box vegetation at the ends of the peninsulas, long-term salt accumulation is minimal (Fig. 25). Local groundwater salinity at the ends of peninsulas is low, as most saline regional groundwater has been discharged as baseflow at much narrower parts of the floodplain. High water use from vegetation has lowered groundwater tables and led to an influx of fresh water into the alluvial aquifer from the river.

Vegetation health mapping (Telfer and Overton, 1999) shows a concentration of dead trees or trees in poor health around the backwater on Clarks Floodplain (Fig. 3), corresponding with areas of high salt accumulation (Fig. 25). Poor vegetation health is also evident in the combined red gum and black box stands around the outside of the bend correlating with higher salt accumulation within divisions 4–7. High rates of salt accumulation indicate areas where vegetation health or water bodies are under threat from salinisation, and focus where remediation and management efforts should be directed. Methods such as direct irrigation during drought, targeted spear-point pumping of groundwater below affected vegetation and flood augmentation may be used to reduce the salt impact on floodplain ecosystems.

Interrelationships between vegetation and groundwater hydrology

The analysis section of this paper has treated vegetation distribution as an independent variable in order to ascertain
causal factors for groundwater flux distribution, when in reality it has complex relationships with geomorphology, groundwater hydrology and salinity. Vegetation with high water requirements such as the red gum eucalypt is more likely to be found in areas with access to fresh groundwater, such as the inside of meanders or other sandy deposits with good connection to the river. High rates of evapotranspiration are able to occur as a result of the available fresh water.

Floodplain width, tree area and amplification factor were found to be highly correlated, with correlation coefficients ranging between 0.93 and 0.96, with wider floodplain divisions having high tree coverage and high amplification due to the presence of meanders. Correlation was less pronounced for floodplain elevation, soil type and highland inflow.

The value of these results lies in determining the most critical parameters to include in groundwater models to adequately represent groundwater discharge processes on a fine spatial scale. In finite difference numerical models, varying the evapotranspiration function spatially according to vegetation distribution will give a more appropriate representation of groundwater flow processes than uniform parameters, or variation of evapotranspiration with soil or other factors. For full ecological analyses, however, the lack of feedback on groundwater salinity from salt accumulation will be a disadvantage for temporal investigations of vegetation health responses using this methodology.

Conclusions

This paper has provided a discussion of the importance of geometry, geomorphology and elevation in determining the spatial distribution of discharge from a floodplain. It used a quasi-three-dimensional MODFLOW model to confirm the critical parameters required to represent groundwater discharge patterns affecting vegetation health, and compared these results with a uniform MODFLOW model using no spatial distribution of vegetation or elevation parameters.

The key findings from this study are that:

- To assess the magnitude of the impact on floodplain biodiversity and ecological value, groundwater modelling needs to take into account spatial patterns of groundwater discharge and rates of salinisation on a fine scale. A uniform model with no spatial variation or recharge information was not able to accurately represent spatial distributions of evapotranspiration.
- Progressively more information on the spatial patterns of discharge was gained when moving from a uniform parameter floodplain model to variation of elevation, to variation of vegetation and soil parameters. Better correlations between model results and field measurements were achieved when vegetation parameters were included in the modelling.
- Due to the large range of vegetation present on the floodplain studied, the distribution of vegetation type had the most significant effect on the spatial distribution and magnitude of groundwater discharge between groundwater flow divisions. Floodplain elevation and width had a lesser effect, but the distribution of vegetation related to both of these parameters.
- Localised elevation is the most critical parameter in determining the proportion of inflow to a floodplain expressed as seepage and is significant in determining spatial patterns of evapotranspiration. Geomorphic features such as backwaters, lagoons and palaeochannel, which are commonly found in meandering river systems, can produce elevation changes that significantly affect the spatial distribution of discharge.
- The complex meandering patterns of the river defined where baseflow occurs on a fine scale. Higher proportions of baseflow are associated with areas of floodplain that are close to the break of slope, and where river frontage, or division amplification is low. Negative baseflow was found on the inside of river bends where eucalypt communities are present, and where division amplification is high. The uniformly parameterised model was able to represent baseflow with some degree of accuracy, but could not predict the negative baseflow associated with high vegetation water use.
- The uniformly parameterised and more complex distributed models produced similar baseflow distribution despite very different spatial patterns of evapotranspiration. While many groundwater models are calibrated or validated against baseflow calculations, further comparison with spatial indicators of groundwater discharge, such as vegetation water use or salt accumulation maps are required when vertical groundwater flux is of interest.

Although long-term patterns of net groundwater discharge were found to depend on vegetation distribution, elevation, soil type and river geometry, the relationship between these interdependent parameters is complex, and based on the geomorphology and ecology of the riverine system. The value of these results, however, lies in determining the most critical parameters to include in groundwater models to adequately represent spatial groundwater discharge processes.

Acknowledgements

The first author acknowledges the support of Land and Water Australia and the Centre for Groundwater Studies. The work was also supported through the project 'Assessing Current and Future Impacts of Land Management Induced Groundwater Discharge on Floodplain Health', which is a partnership between the River Murray Catchment Water Management Board (RMCWMB), CSIRO Land and Water, the South Australian Department of Environment and Heritage, and South Australian Department for Water, Land and Biodiversity Conservation (DWLB). The project was funded through the Natural Heritage Trust and by the project partners, and managed by RMCWMB.

References


