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The impact of flooding on modelling salt transport processes to streams

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Abstract

The development of many of the world's arid and semi-arid regions has resulted in the salinisation of land and water resources. In these areas, soils and groundwaters are often naturally saline and any disturbance of the delicate hydrological balance results in mobilisation of the stored salt. The salt transport mechanisms are often highly complex, the understanding of which necessitates the use of computer modelling in combination with field studies. In this paper the transport of salt between groundwater and streams on the Chowilla floodplain in south-eastern Australia was modelled and compared with available field data. The large salinity contrast between the fresh stream and floodwater and the saline groundwater results in density-dependent flow behaviour, and hence necessitated the use of a variable density flow and solute transport model (SUTRA). The model was applied in cross-section over a 6.1-km-long transect across the floodplain. Time varying boundary conditions were employed at the locations of three streams on the transect to simulate the interaction between the rising and falling streams and the adjacent aquifer during and after floods. The model was used to assess the importance of overbank floods in the transport of salt to floodplain streams by carrying out simulations under various recharge scenarios. The simulations showed that the mixing of floodwater and groundwater within the bank storage adjacent to the streams could predict the observed short-term ($<$ 12 months) salt load recessions. In order to predict the observed long-term (12–24 months) salt load recessions, the inclusion of localised recharge during overbank floods is required, as hypothesised by previous field-based studies.

Keywords: Stream–aquifer interaction; Density-dependent flow; Floodwater; Groundwater; Salt transport; Floodplain; SUTRA

Software availability

Program Title:	SUTRA
Developer:	Dr C. I. Voss
Contact Address:	United States Geological Survey, 431 National Center, Reston, Virginia 20192, USA.
E-mail:	cvoss@usgs.gov
Hardware Required:	IBM PC compatible, UNIX workstation
Program Language:	FORTRAN
Availability and Cost:	Free from http://h2o.usgs.gov/software/ground_water.html

1. Introduction

Approximately one third of the world's land area is either arid or semi-arid (Rogers, 1981). Development of these regions for agricultural production has in many cases resulted in serious environmental consequences, with salinisation of rivers and streams being one of the most prominent (e.g. Orlob and Ghorbanzadeh, 1981; Konikow and Person, 1985; Allison et al., 1990).

The Murray–Darling Basin in south-eastern Australia (Fig. 1) is a notable example of where stream salinisation is widespread and of serious concern. This extensive river system drains an area of approximately 10^6 km^2 (one-seventh of the Australian continent), most of which

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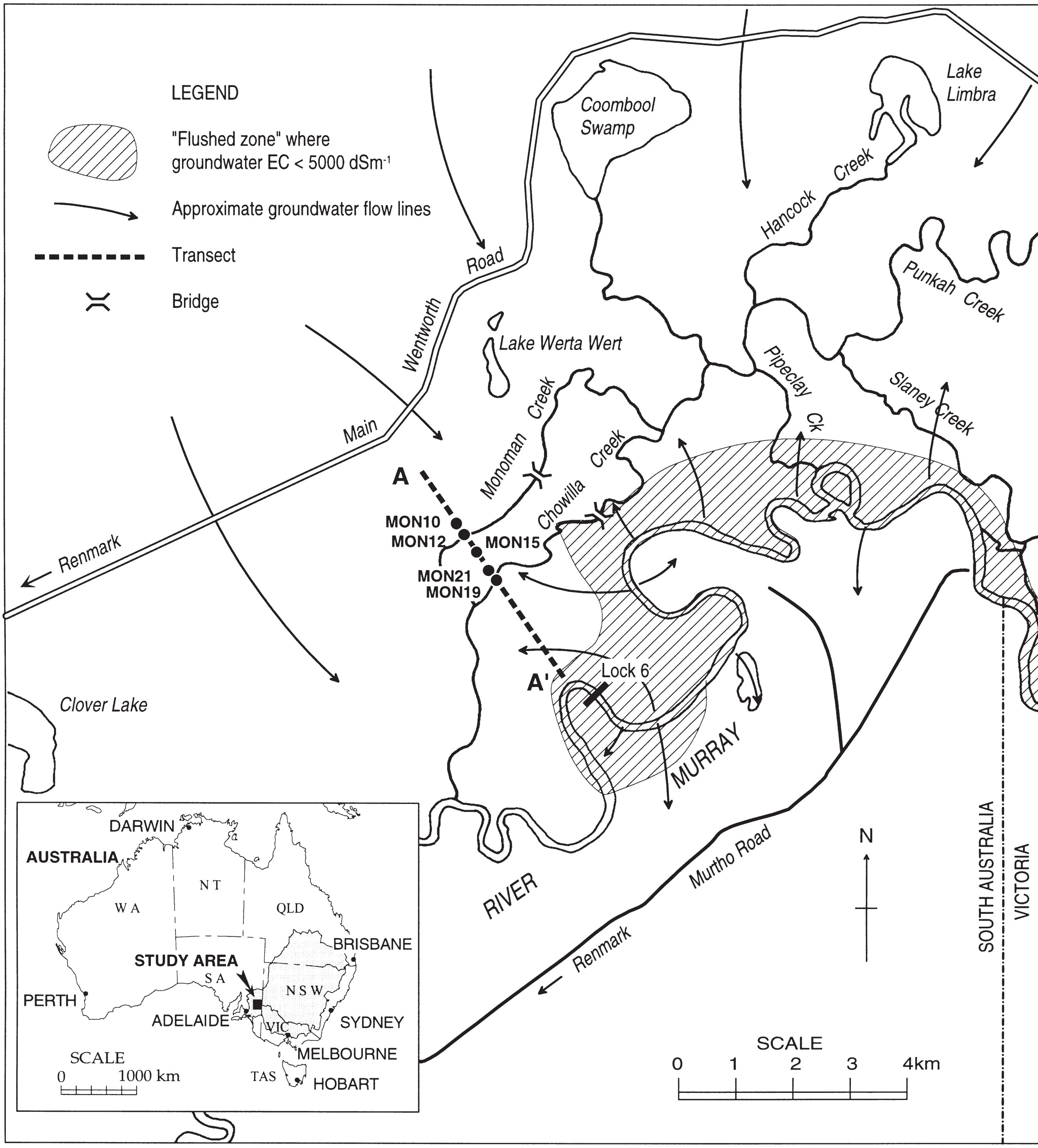

Fig. 1. The Chowilla anabranch in the vicinity of Monoman Island. Groundwater flow directions during non-flood times, location of the "flushed zone", and the transect A–A′ of piezometers used for the modelling are shown. The location of the study site within the Murray–Darling Basin is shown in the inset.

is either arid or semi-arid. The salinity of the main tributary, the River Murray, increases with distance downstream with the largest increases occurring in the lower one third of the system where the median salinity (expressed as Electrical Conductivity, EC) doubles from 400 μS cm^{-1} at the South Australia/Victoria border to 800 μS cm^{-1} at its exit to the sea (Fig. 1). The large increases in this reach are due to the discharge of saline (> 20000 μS cm^{-1}; Evans, 1988) groundwater from regional aquifers to the river, an essentially natural process exacerbated by irrigation activities in close proximity to the river. Of considerable concern is that the salinity of the River Murray in this reach is increasing in time, as illustrated by the study of Morton and Cunningham (1985) which showed that river salinity had increased by > 2% per year since 1970.

The flow of saline groundwater to the lower reaches of the River Murray is complicated by the presence of floodplains up to 10 km in width. In these areas, the depth to groundwater is generally less than 4 m, leading to significant loss of groundwater by evapotranspiration (Jolly et al., 1993). Periodic flooding displaces additional salt into the river, either directly or via floodplain streams, causing a dramatic increase in salt loads to the river, with the elevated salt accessions continuing for several months after a major flood (Fig. 2). The processes which lead to this behaviour are poorly understood but are postulated to be due to localised floodwater recharge on the floodplain (Jolly et al., 1994). This hypothesis is difficult to confirm in the field and so a modelling approach is necessary.

The salinisation process is governed by the nature of the stream–aquifer interaction. When the river has an adjacent floodplain, stream–aquifer interactions can be highly complex due to the influence of overbank floods. This complexity dictates that field studies alone are often insufficient to unravel the nature of the underlying processes operating in a given situation. Moreover, they are of only limited use in determining the probable impacts of changed river and/or floodplain management. For these reasons groundwater flow models, both analytical (e.g. Hall and Moench, 1972; Morel-Seytoux and Daly, 1975) and numerical (e.g. Marino, 1975, 1981), have been developed for the study of stream–aquifer interaction. While many of the models account (in various ways) for the time varying stream boundary conditions, they are limited in that they only consider the flow of water into and out of bank storage and do not model the effects of overbank flow which occurs during floods. Furthermore, they are concerned only with the transport of water between the stream and the aquifer and not that of any solute. Notable exceptions to the latter are the studies of Konikow and Bredehoeft (1974); Konikow and Person (1985); Person and Konikow (1986).

A phenomenon often typical of stream–aquifer interaction in arid and semi-arid areas is the large contrast in salinity between the fresh stream/floodwater and groundwater. The standard approach to modelling salt transport to streams is to utilise a groundwater flow model to predict the water fluxes to the stream and to multiply these by the observed or inferred salt concentrations in the aquifer (e.g. Ghassemi et al., 1989). However, studies such as those of Herbert et al. (1988); Oostrom et al. (1992) suggest that large salinity contrasts can

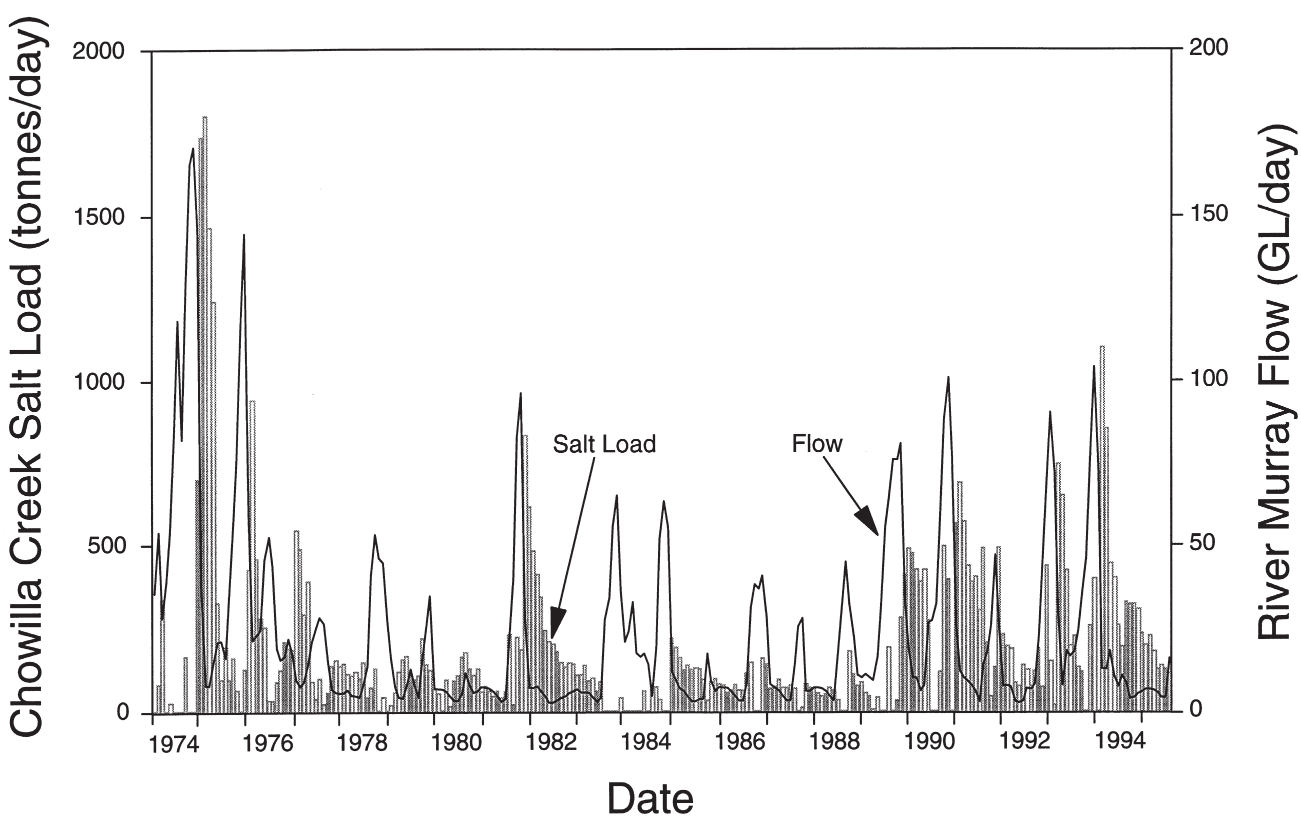

Fig. 2. River Murray flow (solid lines, in GL day^{-1}) and salt load from the Chowilla floodplain region (bars) for the period 1974–1994.

result in density-dependent flow, a phenomenon rarely considered in studies of stream–aquifer interaction.

The aim of this paper is to use a variable density flow and solute transport model (SUTRA) to better understand the processes operating in the transport of saline groundwater to floodplain streams. The SUTRA model (Voss, 1984) has been previously used for numerical simulation of density-dependent flow (Voss and Souza, 1987; Souza and Voss, 1987; Ghassemi et al., 1990; Narayan and Armstrong, 1995). Besides density effects, the model can be modified to handle time varying boundary conditions typical of stream–aquifer interaction. The model is used to simulate stream–aquifer interaction in order to determine the impact of various flooding scenarios on the movement of salt to floodplain streams. Time varying boundary conditions are specifically used to model the rise and fall in river and stream levels and the model is also run in the saturated/unsaturated mode. The simulated results are compared with available field data. Our intention is not to exactly replicate the salt loads observed during and after floods (the paucity of the field data does not allow this comparison) but to simulate the generally observed behaviour.

2. Site description

2.1. Study area and hydrology

The site selected for study was the Chowilla anabranch, a 200 km^2 floodplain situated near the junction of the South Australian, Victorian and New South Wales borders (Fig. 1). The Chowilla anabranch consists of a network of streams which flow from the River Murray upstream of Lock 6 (one of a number of flow control weirs on the lower River Murray) and eventually join together to form Chowilla Creek which discharges back into the River Murray downstream of Lock 6. The region in the vicinity of Monoman and Chowilla Creeks, and Lock 6 is described in this study.

The Chowilla region has a semi-arid climate characterised by mild winters and long hot summers. Mean annual rainfall is about 260 mm and potential evaporation is about 2000 mm. Rainfall is highly variable both within and between the seasons with extended dry spells being common. The distribution of mean monthly rainfall has a slight winter dominance.

Prior to the commencement of flow regulation in the 1920–1930s, the River Murray had a varying flow regime. Large floods occurred in spring and early summer and at other times the river receded to little more than a series of saline waterholes. At most times the river acted as a drain, intercepting regional saline groundwater. River regulation has increased the return period of low to medium size floods by a factor of approximately three (Ohlmeyer, 1991).

The streams of the Chowilla anabranch were ephemeral and flowed only during the periods of flood prior to the construction of Lock 6. The lock creates a permanent upstream weir pool over 70 km long which drives water continually through the Chowilla floodplain streams (40–80% of the River Murray flow is now diverted through the continually flowing creek system). The streams intercept the saline regional groundwater which flows into the region (Fig. 1). At non-flood times baseflow to the creek system contributes a salt load of about 40 t day^{-1} to the River Murray (NEC, 1988). Moreover, inundation of the floodplain causes large volumes of saline groundwater to flow into the creeks following a major flood resulting in salt loads as high as 1800 t day^{-1} (NEC, 1988). Elevated salt loads often continue for up to two years after a flood event (Fig. 2), with the resulting increases in River Murray salinities being detrimental to downstream water users.

2.2. Hydrogeology

The Chowilla region acts as a natural sink for the regional groundwater systems. An overview of the hydrogeology of the Chowilla region is presented in Waterhouse (1989) and a generalised cross-section of the floodplain is shown in Fig. 3. The main aquifers are the Monoman Formation and the Pliocene Sands aquifer.

The Monoman Formation underlies the floodplain and consists of alluvial sands approximately 30 m in thickness. The water table formerly resided in this aquifer, but due to the construction of Lock 6 it has risen 2–3 m (MDBC, 1995) and is now often found within the overlying Coonambidgal Formation causing the aquifer to be semi-confined. The salinity of this aquifer ranges from 20 000 to 60 000 mg L^{-1} total dissolved solids (TDS) with the highest values at the base of the aquifer (Jolly et al., 1992). However, the weir pool upstream of Lock 6 has enhanced groundwater recharge in the immediate vicinity and created a zone of low salinity ($<$ 3000 mg L^{-1} TDS) groundwater known locally as the "flushed zone" (Fig. 1). The Monoman Formation is in direct hydraulic contact with the Pliocene Sands aquifer and also has good hydraulic connection to the creeks and the River Murray. The horizontal hydraulic conductivity of this aquifer is in the range 4–10 m day^{-1}.

The Pliocene Sands aquifer consists of estuarine fine, medium and coarse sand with some more clayey layers. The clay content generally increases towards the base of the aquifer. The aquifer is unconfined and has salinities in the range 40 000–90 000 mg L^{-1} TDS. The horizontal hydraulic conductivity is lower than that of the Monoman Formation and is of the order of about 5 m day^{-1}. Regional groundwater flow occurs laterally from the Pli-

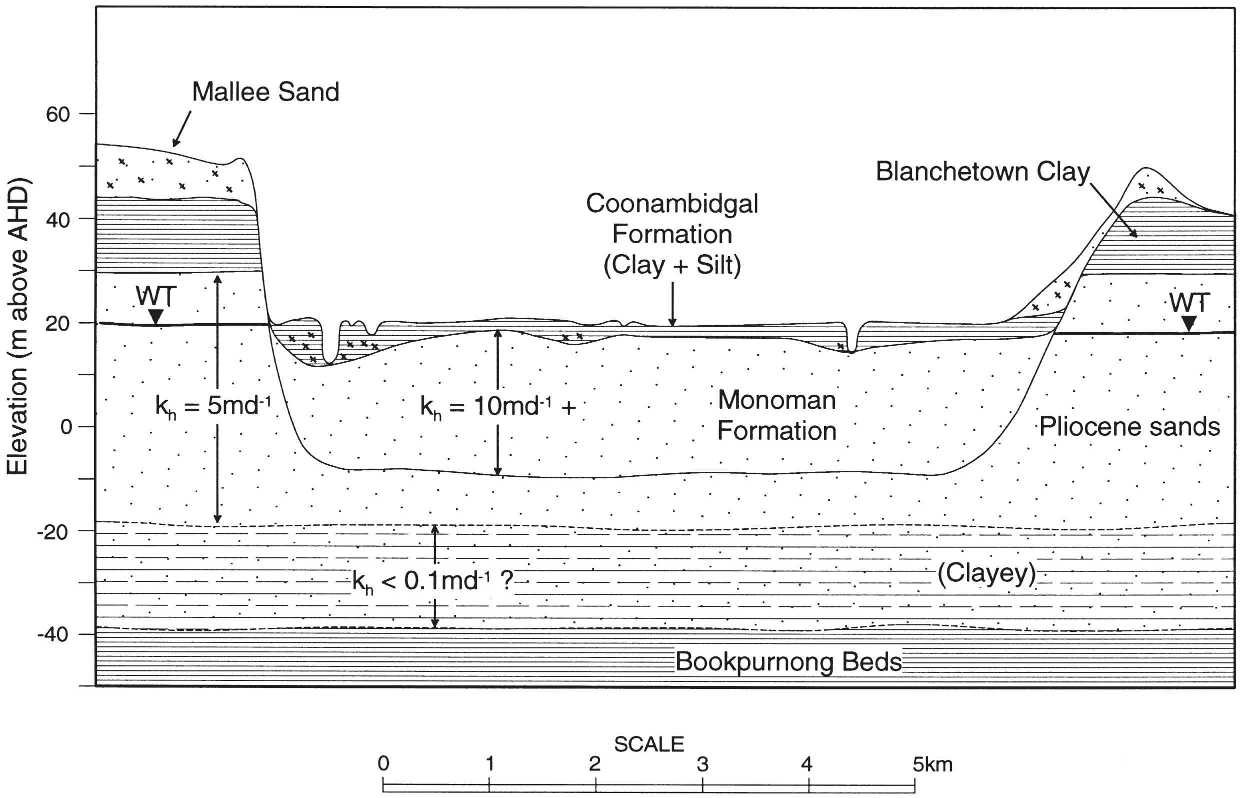

Fig. 3. Generalised hydrogeological cross-section of the Chowilla floodplain (after Waterhouse, 1989). Vertical elevation is given in metres above the Australian Height Datum (AHD).

ocene Sands aquifer towards the floodplain, entering the Monoman Formation from the north.

2.3. Soils

The soils of the Chowilla floodplain have been studied in detail by Hollingsworth et al. (1990). In brief, the soils of Chowilla region generally consist of a micaceous cracking clay of alluvial origin (Coonambidgal Formation) which has minimal profile development except for occasional gypsum, halite or calcium carbonate horizons. The Coonambidgal Formation can be up to 5 m in thickness, with the deepest deposits occurring close to existing or prior creeks. In some locations (most notably the higher elevation land units, and in limited areas, the beds of the floodplain creeks) the Coonambidgal Formation is absent. As shown by Jolly et al. (1994) the dispersive nature of this soil results in very low hydraulic conductivities, especially when inundated by low salinity floodwaters. The Coonambidgal Formation overlies the Monoman Formation.

3. Methods

3.1. Field measurements

3.1.1. Piezometric levels and chemistry

A network of piezometers was installed to supplement the limited number of existing monitoring bores at the site. These were installed at various depths within the Monoman Formation along the transect (A–A′) shown in Fig. 1, and at other locations in the vicinity of Monoman Island. Water levels in these piezometers were monitored throughout the course of a flood in 1990/1991 and they were sampled for chemistry prior to and following the flood. Details of the water level monitoring, water quality sampling, and results are presented in Jolly et al. (1992). In brief, the piezometers at all depths in the aquifer rose and fell in response to the flood, with the most marked responses occurring at sites close to the creeks. Water quality within the aquifer generally did not change due to the flood. Notable exceptions were some of the shallow piezometers located close to the creeks where salinities initially fell as a result of the flood, but rose again in the following months.

3.1.2. River/creek levels and salinity

A number of sources of information were used in describing the water levels of Monoman and Chowilla Creeks. Daily levels of the upper and lower pools of Lock 6 were obtained from the Engineering and Water Supply Department of South Australia (EWSD) for the period 1970–1992. Jolly et al. (1994) showed that during times of flood there was excellent correlation between Lock 6 lower pool level and the levels of Monoman and Chowilla Creeks.

The levels of Monoman and Chowilla Creeks were measured on a weekly basis by the EWSD during the period January 1989 to April 1990. These measurements were made at the Monoman and Chowilla Creek bridges, respectively (Fig. 1). Concurrently, the salinity (as electrical conductivity) at the centre of each creek was measured at the top and bottom of the water column. Both the water level and salinity records are incomplete as no measurements were taken during the 1989 flood (July–November). In spite of this, these data provide useful information during the recession of the 1989 flood, as will be shown later.

Salt loads in Monoman and Chowilla Creeks were obtained from data presented in MDBC (1995). In the period 1972–1991 the EWSD carried out several studies of the surface hydrology of the Chowilla system. In each study creek flow and salinity was measured at selected locations allowing calculations to be made of salt loads from various reaches. These data were used to estimate the approximate salt loads (in t day^{-1} km^{-1}) in Chowilla and Monoman Creeks (in the vicinity of Monoman Island) at 12 times during the period 1976–1986. These data, while of some use in determining the non-flood salt loads to both creeks, are insufficient to construct salt load curves for various size floods. However, salt loads from the entire system have been measured on a monthly basis since 1974 (Fig. 2). These data were synthesised in MDBC (1995) to produce "characteristic" salt recession curves for floods with peak flows of about 60 000 ML day^{-1} and 100 000 ML day^{-1} respectively, and these are used for comparison with the model results (Fig. 4).

3.2. Numerical modelling

3.2.1. SUTRA model

Due to the contrast in salinities of the groundwater (often greater than seawater) and the creek and floodwaters (fresh), density-induced flow was thought to be important in the transport of salt to the River Murray. The density-dependent model SUTRA (*S*aturated–*U*nsaturated *Tra*nsport) was used to simulate changes in the aquifer pressure (and hence head) and salt concentration in response to flooding. SUTRA is a numerical model which simulates fluid movement and the transport of either energy or dissolved substances in a subsurface environment (Voss, 1984). The model employs a two-dimensional finite element approximation of the governing equations in space and an implicit finite difference approximation in time. For solute transport, two partial equations are solved simultaneously (fluid and salt mass-balance equations). The model is stable and accurate when employed with proper spatial and temporal discretisation. SUTRA computes the pressure and solute concentration at each node for each time step. In order to compare the predictions of pressure with field measurements of piezometric head, the pressures can be transformed to equivalent freshwater heads. Although the use of freshwater heads is not strictly valid when considering variable density flows (Davies, 1987), it was felt appropriate for comparison with the field measured piezometric heads.

3.2.2. Model conceptualisation, spatial discretisation and boundary conditions

In order to model salt transport to Monoman and Chowilla Creeks a 1-m-thick vertical slice was taken along transect A–A′ (Fig. 1). Several piezometers which lie along the transect were used in the model calibration (MON10, MON12, MON15, MON19 and MON21; Fig. 1). To represent the major features of the system the mesh was discretised in a non-uniform manner with quadrilateral elements consisting of 2914 (31 × 94) nodes and 2790 elements (Fig. 5). The mesh extends 6100 m in the horizontal direction and 46 m in the vertical direction. For spatial discretisation, a fine mesh spacing of 0.5 m to the sides of the creeks allowed accurate modelling of the clay layers around the creeks. The vertical spacing was also non-uniform varying from 0.5 m around the base of the creeks to 5 m at the base of the aquifer. The mesh was generated using FEMCAD software developed by Beer and Mertz (1990) and then converted to the required SUTRA format using an interface package developed by Buia et al. (1994).

The boundary conditions for the simulation are also shown schematically in Fig. 5. The near surface Coonambidgal Formation which confines the aquifer is simulated as the top model boundary. The bottom of the mesh is located at the boundary between the permeable Pliocene Sands aquifer and its low permeability clayey lower section. This region and the underlying Bookpurnong Beds are considered relatively impermeable and so a no flow boundary is utilised at the bottom of the mesh. On the southern boundary (River Murray) the river depth is assumed to be 11 m where a time dependent boundary condition is employed to simulate the rising and falling heads during the flood. A no flow boundary condition is employed below the bed of the River Murray. A specified head boundary of 17 m AHD (Australian Height Datum) is imposed along the northern vertical boundary (edge of floodplain). Monoman Creek is located 1560 m from the northern edge of the floodplain and Chowilla

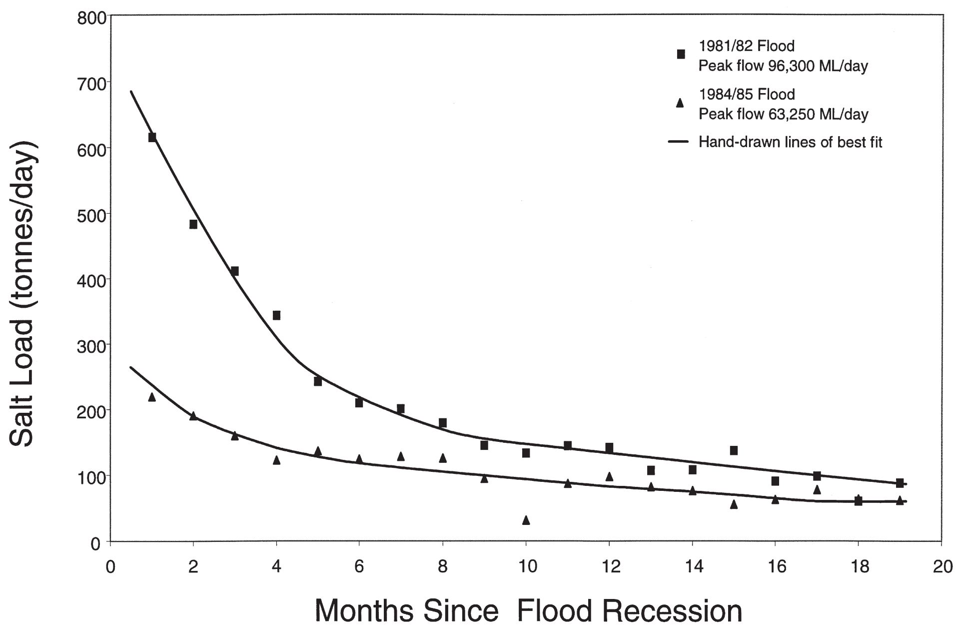

Fig. 4. Salt load recessions for the 1981/1982 and 1984/1985 floods for the entire Chowilla anabranch system. The solid lines are hand-fitted approximations of the salt load recession curves (after MDBC, 1995).

Creek is located at a distance of 2660 m. Time dependent pressure boundary conditions are imposed on both creeks. Groundwater inflow occurring through the northern specified head boundary has the concentration of 45 000 mg L^{-1} TDS and inflow from the River Murray and both creeks has a concentration of 350 mg L^{-1} TDS. Any flow out of the mesh at the specified head boundaries occurs at the ambient concentration of the aquifer fluid. Solute may neither disperse nor advect across no-flow boundaries.

3.2.3. Model parameters

Values used for input into the SUTRA model are given in Table 1. The hydraulic conductivities given by Waterhouse (1989) of the upper part of the Monoman Formation are 10–12 m day^{-1}, the lower part of the Monoman Formation are 3–4 m day^{-1} and the Pliocene Sands are 3–6 m day^{-1}. For simplicity, the upper Monoman Formation was simulated with a value of 10 m day^{-1} and both the lower Monoman Formation and the Pliocene Sands were simulated with a value of 4 m day^{-1}. The ratio of vertical hydraulic conductivity to horizontal conductivity in all aquifers was taken to be 1:10 (MDBC, 1992). The porosity of both aquifers was assumed to be 0.3.

Field scale dispersion has not been measured for any of the aquifers and thus estimates of longitudinal dispersivity (α_L) has been primarily based on guidelines provided by Voss (1984) for proper discretisation. In most cases, to guarantee spatial stability the condition $\Delta_L \leq 4\alpha_L$, must be enforced where Δ_L is the local distance between element sides along a flow line. Given the maximum horizontal element length is 205 m, it was considered appropriate to take $\alpha_L = 20$ m, a value within the range of data for sand given in Gelhar et al. (1992). Transverse dispersivity (α_T) is normally lower by a factor of 5–20 and so a value of $\alpha_T = 1$ m was used.

The saturated/unsaturated mode of SUTRA was applied using the three parameter equations of Van Genuchten (1980) which relate capillary pressure (p_c) to percentage saturation (S_w) and relative hydraulic conductivity (k_r):

$$S_w = S_{wres} + (1 - S_{wres})\left[\frac{1}{1 + (ap_c)^n}\right]^{\left(\frac{n-1}{n}\right)} \tag{1}$$

$$k_r = \left(\frac{S_w - S_{wres}}{1 - S_{wres}}\right)^{\frac{1}{2}}\left\{1 - \left[1 - \left(\frac{S_w - S_{wres}}{1 - S_{wres}}\right)^{\left(\frac{n}{n-1}\right)}\right]^{\left(\frac{n-1}{n}\right)}\right\}^2 \tag{2}$$

where S_{wres} is residual saturation (dimensionless), a is a fitting parameter (m s^{-2} kg^{-1}) and n is a fitting parameter (dimensionless). Note that S_{wres}, a and n in Eqs (1) and (2) (and in Voss, 1984) correspond to θ_r, α and n in Van Genuchten (1980). Based on previous work at the study site by Jolly et al. (1993), values typical of a silty clay loam were used for the Van Genuchten (1980) parameters (Table 1). These parameters were fixed during calibration as the intention was not to simulate the detailed behaviour of the unsaturated zone, but rather to permit rapid saturation and de-watering of the materials in the banks of the two streams as water levels in the streams changed with time.

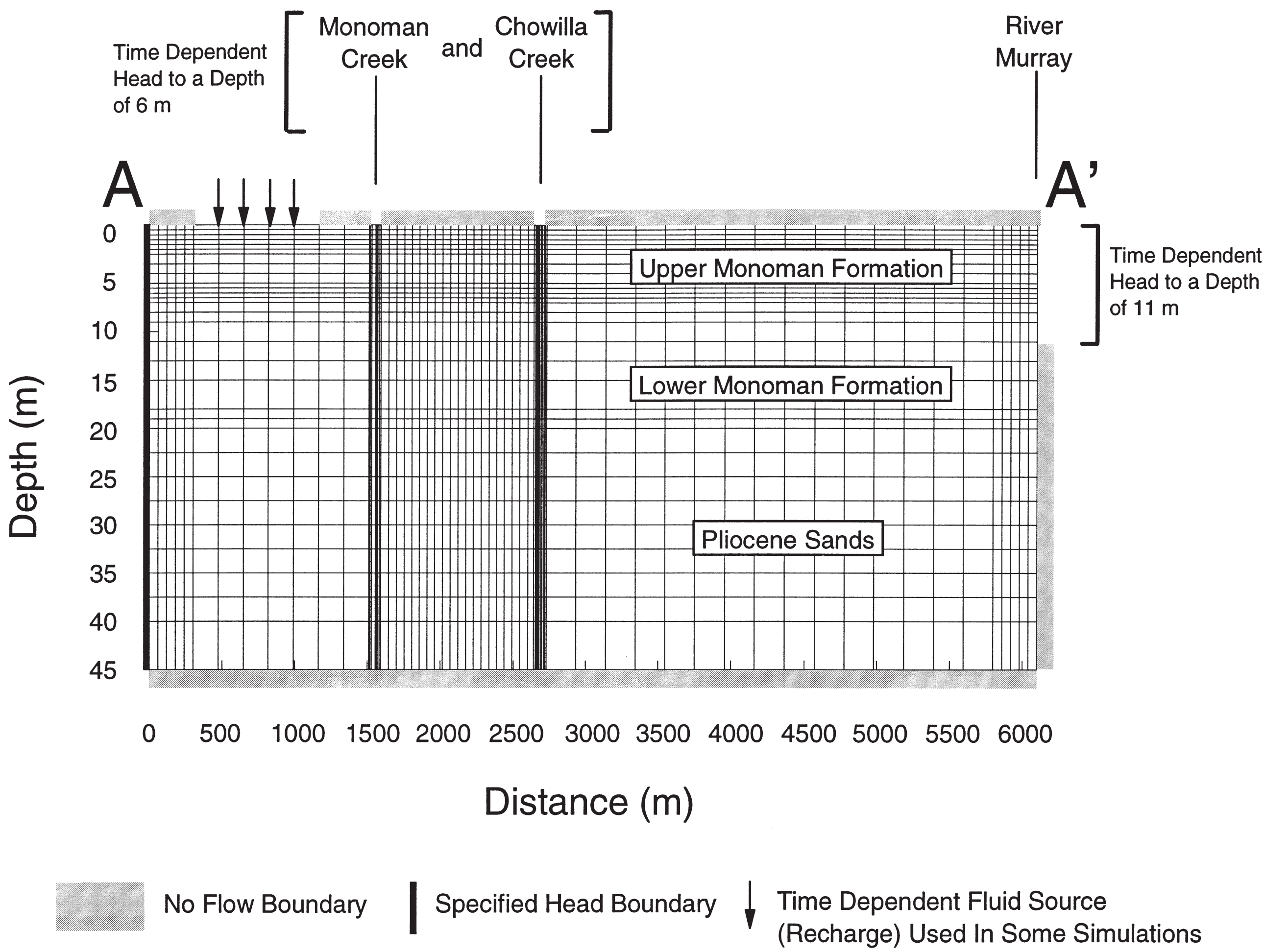

Fig. 5. Model mesh and boundary conditions of the simulated cross-section A–A′. The boundaries between the Upper and Lower Monoman Formations and between the Lower Monoman Formation and the Pliocene Sands are located at depths of 6 and 26 m, respectively.

3.2.4. Time dependent boundary conditions

The River Murray boundary condition was represented by the upper pool of Lock 6 and the Chowilla and Monoman Creek boundary conditions were represented by the lower pool of Lock 6. Functional approximations of the time-dependent boundary conditions during flooding were incorporated in the BCTIME subroutine of SUTRA using:

$$H(t) = H_{\min} + \frac{H_{\max} - H_{\min}}{2}\left[a\left[1 + \sin\left(\left(\frac{2t}{T} + \frac{3}{2}\right)\pi\right)\right]\right]^{b} \tag{3}$$

where $H(t)$ is a variable head function, $H_{\max}$ and $H_{\min}$ are maximum and minimum head levels (m), $t = n\Delta t$ is simulated time (days), where n is an integer and Δt is a time step size used in the simulation (days), T is the period of fluctuation (days) and a and b are curve fitting constants.

Different forms of Eq. (3) were used to represent the rising and falling levels of the River Murray and the two creeks, and each function was calibrated by trial and error to represent a flood with peak flows of 100 000 ML day^{-1}. Since Chowilla and Monoman Creeks are hydraulically connected the same function was used for each creek. The values of a, b, T, $H_{\min}$ and $H_{\max}$ used for the River Murray and both creeks are given in Table 2. As shown in Fig. 6, the two functions are well matched with field data from the 1990/1991 flood (which had a peak flow of 101 900 ML day^{-1}).

3.2.5. Initial conditions and time discretisation

In order to produce realistic initial pressure and concentration conditions for the flooding scenarios it was necessary to carry out two preliminary simulations. The first was a steady-state non-flood run to synthesise conditions prior to the installation of Lock 6. The boundary conditions for this simulation were the River Murray head fixed at a height of 17 m AHD and a groundwater salinity of 350 mg L^{-1} TDS, and a constant head boundary of 19 m AHD at the northern end at a groundwater

Table 1
Model parameters

Parameter	Value and unit
Freshwater density (ρ)	1000 kg m^{-3}
Groundwater density at salinity of 45,000 mg L^{-1} TDS (ρ_{gw})	1030 kg m^{-3}
Fluid dynamic viscosity (μ)	10^{-3} kg m^{-1} s^{-1}
Coefficient of fluid density change ($\delta\rho/\delta C$)	700 kg m^{-3}
Water compressibility (β)	4.5×10^{-10} Pa^{-1}
Soil compressibility (α)	3×10^{-8} Pa^{-1}
Porosity of Monoman Formation and Pliocene Sands (ϵ)	0.3
Porosity of clay lining around Chowilla Creek (ϵ)	0.3
Porosity of clay lining around Monoman Creek (ϵ)	0.6
Specific pressure storativity of Monoman Formation and Pliocene Sands	2.1×10^{-8} Pa^{-1}
Specific pressure storativity of clay lining around Chowilla Creek	2.1×10^{-8} Pa^{-1}
Specific pressure storativity of clay lining around Monoman Creek	1.2×10^{-8} Pa^{-1}
Hydraulic conductivity of upper Monoman Formation (K_h)	10 m day^{-1}
Hydraulic conductivity of lower Monoman Formation (K_h)	4 m day^{-1}
Hydraulic conductivity of Pliocene Sands (K_h)	4 m day^{-1}
Hydraulic conductivity of clay lining around Chowilla Creek	6×10^{-2} m day^{-1}
Hydraulic conductivity of clay lining around Monoman Creek	6×10^{-4} m day^{-1}
Ratio of vert. to horiz. hydraulic conductivity for all aquifers (K_v/K_h)	0.1
Longitudinal dispersivity for all aquifers (α_L)	20 m
Transverse dispersivity for all aquifers (α_T)	1 m
Molecular diffusivity for all aquifers (D_m)	1.5×10^{-9} m^2 s^{-1}
Residual saturation (S_{wres})	0.1
Van Genuchten (1980) parameter (a)	5×10^{-5} m s^{-2} kg^{-1}
Van Genuchten (1980) parameter (n)	2.0
Upper Monoman Formation thickness	6 m
Lower Monoman Formation thickness	20 m
Pliocene Sands thickness	20 m
River Murray depth	11 m
Monoman Creek depth	6 m
Chowilla Creek depth	6 m
Length of model domain	6100 m
Depth of model domain	46 m
Thickness of model domain	1 m

Table 2
Parameters used for the functional approximations of the time-dependent boundary conditions (Eq. (3)) for a flood with a peak flow of 100,000 ML day^{-1}

Boundary	t (days)	a	b	T (days)	H_{min} (m AHD)[a]	H_{max} (m AHD)[a]
River Murray	1–163	0.54	10.0	362	19.25	20.35
	164–200	0.71	4.0	378	19.80	20.10
	201–400	0.54	16.0	362	19.25	20.35
	401–800	–	–	–	19.25	19.25
Creeks	1–200	0.95	1.1	362	16.30	20.35
	201–400	0.58	5.5	362	16.30	20.35
	401–800	–	–	–	16.30	16.30

[a]Metres above Australian Height Datum.

salinity of 45 000 mg L^{-1} TDS. Both Monoman and Chowilla Creeks were not simulated as the floodplain creeks did not flow during non-flood times prior to the installation of Lock 6. Using the results of this simulation, a 60 year transient run was carried out with the River Murray head raised to its present non-flood level of 19.25 m AHD, and the heads of both Monoman and Chowilla Creeks set at their present level of 16.3 m AHD. The results of this run were used as the initial conditions for the flooding scenarios.

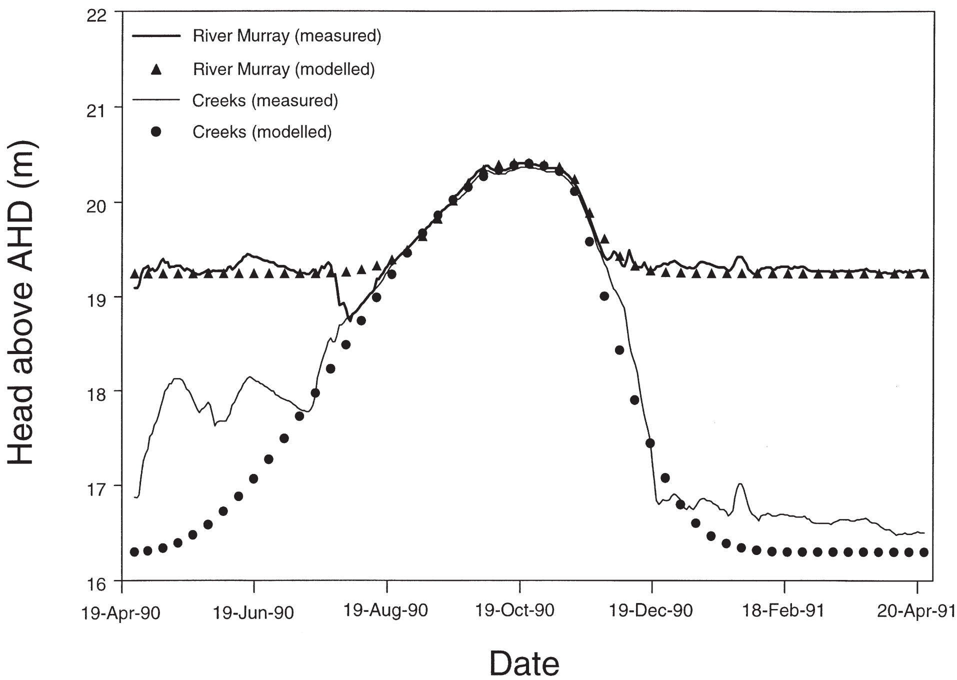

Fig. 6. River Murray hydrographs (Lock 6 upper and lower pools) and sinusoidal functions used to simulate the rise and fall of river and creek heads during the 1990/1991 flood (peak flow of 101,900 ML day^{-1}). The heads are plotted in metres above the Australian Height Datum.

Voss and Souza (1987) discuss the selection of time steps for an accurate numerical solution with changing heads. Each of the flooding simulations were run for 4000 time steps, each of 0.2 of a day duration, giving a total simulation time of 800 days.

4. Results

4.1. Stream salinity measurements

Shown in Fig. 4 are salt load recessions for floods in 1981/1982 (peak flow of 96 300 ML day^{-1}) and 1984/85 (peak flow 63 250 ML day^{-1}) for the entire Chowilla anabranch system. These were derived from the monthly salt load data presented in Fig. 2 and are floods which were not followed the following year by another large flood, thus preserving the characteristics of the salt recession. These are taken to be typical of the salt recessions following floods with peak flows of 60 000 ML day^{-1} and 100 000 ML day^{-1} and are used for comparison with the results of the SUTRA simulations.

Shown in Table 3 are Monoman and Chowilla Creek flows and salt loads at a number of selected times in the period 1976–1986. Also shown for comparison is the River Murray flow at each time. The flow data are from measurements reported in MDBC (1995). The salt loads were estimated by multiplying the flow by the salinity (as electrical conductivity multiplied by a conversion factor of 0.6) difference between two sites located at the upstream and downstream ends of a reach. The salt load estimates should be considered as approximate only with the Chowilla Creek data the least accurate due to the electrical conductivity differences often being very small ($< 5\ \mu S\ cm^{-1}$).

While approximate only, the data in Table 3 provide an indication of the "steady-state" salt loads to both creeks. As there was no flood in 1985, the estimates in February to April 1986 are a good indication of non-flood salt loads. For Chowilla Creek, they were in the range 0.3–1.5 t day^{-1} km^{-1}, while for Monoman Creek the salt loads were 0.2–0.4 t day^{-1} km^{-1}.

The manner in which salt enters the creeks following a flood is illustrated by a plot of the measured salinity at the top and bottom of Monoman Creek and the water level of Chowilla Creek (comparable to that of Monoman Creek) during and after the 1989/1990 flood (Fig. 7). These data show that during the flood the salinities at the top and bottom of the Monoman Creek water column were similar. Immediately following the flood recession, the salinity at the bottom of Monoman Creek increased dramatically as dense saline groundwater began to flow into the creek. While there were increases in salinity at the top of Monoman Creek after the flood, these were much less dramatic than those at the bottom.

4.2. Numerical simulations of groundwater flow and salt transport

4.2.1. Scope of the simulations

Calibration of the numerical model was carried out in a number of steps. This was necessary because the nature

Table 3
Summary of Monoman and Chowilla Creek flows and salt loads measured at selected times in the period 1976–1986

Date	River Murray flow (ML day^{-1})[a,b]	Monoman Creek flow (ML day^{-1})[b]	Monoman Creek salt load (t day^{-1} km^{-1})	Chowilla Creek flow (ML day^{-1})[b]	Chowilla Creek salt load (t day^{-1} km^{-1})
29 Jan. 1976	20 880	450	5.5	3600	12.2
10 Feb. 1976	20 580	300	14.0	2700	0.0
9 Aug. 1976	21 740	350	1.9	2700	1.1
28 Feb. 1977	8680	20	2.5	1030	2.7
25 Jan. 1983	5870	20	0.5	–	–
26 Feb. 1986	6700	50	0.2	1660	0.7
11 Mar. 1986	6200	–	–	1770	1.4
13 Mar. 1986	6040	50	0.3	1820	1.1
14 Mar. 1986	6200	40	0.4	–	–
21 Apr. 1986	4670	30	0.4	1480	0.6
22 Apr. 1986	4670	–	–	1660	0.3
23 Apr. 1986	4140	30	0.4	1640	1.3

[a]Measured at gauging station 426200 which is located approximately 20 km upstream of the Chowilla anabranch.
[b]Megalitres per day.

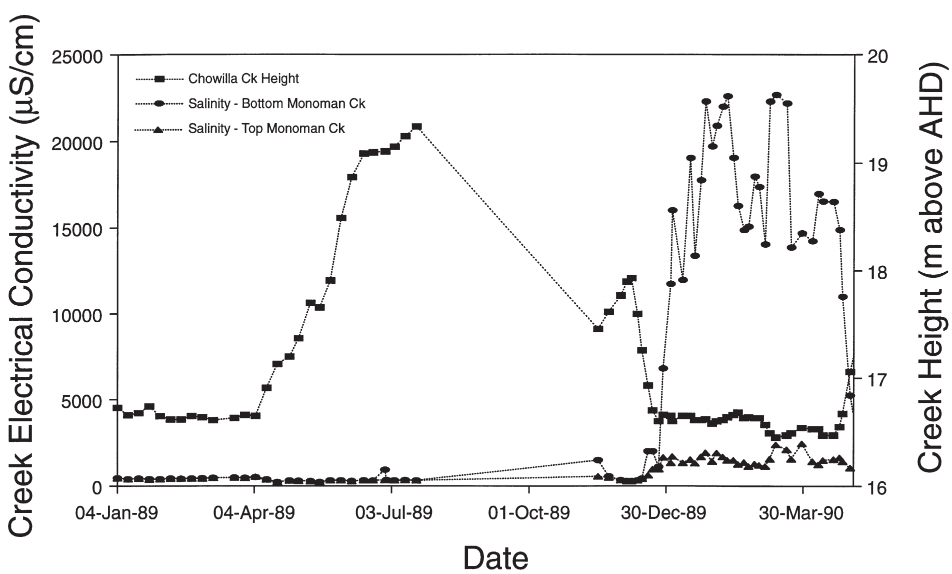

Fig. 7. Chowilla Creek water level (in metres above the Australian Height Datum) and measured salinity (as electrical conductivity) at the top and bottom of Monoman Creek during the 1989/1990 flood (EWSD, unpublished data).

of the hydraulic connection between Monoman and Chowilla Creeks with the aquifer was initially unclear. Calibration was carried out by varying the hydraulic conductivity of the clay lining around Monoman Creek and comparing the model predictions with piezometric data at a number of locations on the model transect (A–A′) during the 1990/1991 flood. In all of the calibration runs the diffuse recharge of floodwater was set to zero. Using the parameters derived from the calibration, comparisons of salt loads and concentration profiles were made with further field data and a number of "what if" simulations were carried out which included the effects of localised recharge on the floodplain.

4.2.2. *Model calibration*

An important step in the calibration process was the determination of the influence of the clay lining of the beds of both Monoman and Chowilla Creeks on the creek/aquifer interaction. During piezometer installation it was observed that the clay layer around Monoman Creek was of much greater thickness and lower hydraulic conductivity than that around Chowilla Creek. The estimated clay thickness around Monoman Creek was 2.5 m at the edges and 1.5 m at the base of the creek. For Chowilla Creek the thickness was estimated to be 0.5 m at both the edges and the base. A porosity of 0.6 was assumed for the clay lining. A number of simulations of the 1990/1991 flood were carried out with the hydraulic conductivity of the Chowilla Creek clay lining fixed at 6×10^{-2} m day^{-1} and that around Monoman Creek varying between 6×10^{-4} and 1.2×10^{-2} m day^{-1}. Comparison of the model predictions with measured data were made at three piezometers along the transect (Table 4). All three piezometers provided reasonable matches to the magnitude and timing of the peak of the hydrograph (Fig. 8). However, the shape of the hydrograph is not well modelled, although the slightly asymmetric rise and fall is simulated. From these and the results for other piezometers (data not shown), a hydraulic conductivity of 6×10^{-4} m day^{-1} was chosen as being appropriate for the clay lining around Monoman Creek.

4.2.3. Predicted and measured piezometric heads, salt concentrations and salt loads during the 1990/91 flood

Using the parameters determined above the model was run to simulate the 1990/1991 flood assuming no diffuse recharge. Shown in Fig. 8 is a comparison of the predicted and measured heads at one of the piezometers close to Monoman Creek (MON10), one in the centre of Monoman Island (MON15) and one close to Chowilla Creek (MON21). Reasonable agreement exists between the measured and predicted heads. Also shown in Fig. 8 are the predicted concentrations at nodes representing piezometers MON 10, MON15 and MON21. While the predicted concentrations do not exactly replicate those measured over the course of the flood (see Jolly et al., 1994), the simulation does predict the general behaviour which was observed; i.e some freshening of the groundwater in bank storage by flood recharge around both creeks, and none at all in the centre of Monoman Island.

The manner in which the concentrations in the piezometers close to creeks rise again after the flood peak are indicative of mixing processes which can explain the short-term ($<$ 6 months) salt loads entering the creeks. Fig. 9 shows modelled contours of concentration around Chowilla and Monoman Creeks at timestep 600 (120 days) which falls on the rising limb of the flood hydrograph. Both creeks show a degree of freshening due to influent creek water in the upper 8 m. Velocity vectors for time steps 600 (120 days) and 1400 (280 days) are given in Fig. 10 for the zone close to Monoman Creek (similar velocity fields also occurred around Chowilla Creek). The direction of flow is clearly from the creek into the top 6 to 8 m of the aquifer at $t = 120$ days and out of the aquifer into the creek at $t = 280$ days (which is at the end of the flood period). It is interesting to note that the highest velocities into and out of the creeks occur at or near their base. This is consistent with the field observations presented in Fig. 7 which show that following flood recession the salinity at the base of the creeks is much higher than at the top of the water column, suggesting that this zone receives the largest volumes of inflowing saline groundwater.

Table 4
Location and depth of screen of piezometers used for comparison with the model predictions

Piezometer	Depth of screen (m)	Distance from nearest creek (m)
MON10	4	4 m from Monoman Creek
MON15	10	Middle of Monoman Island
MON21	10	10 m from Chowilla Creek

Shown in Fig. 11 are the predicted salt loads entering Monoman and Chowilla Creeks during the course of the 1990/1991 flood. Also shown for comparison are predicted salt loads for a much smaller flood (peak flow 60 000 ML day^{-1}). It is interesting to note that in both cases the salt loads to Monoman Creek always exceed those to Chowilla Creek. This has sometimes been observed in the field but is not always the case (Table 3) and appears to confirm the long-term salinity and flow measurements presented in MDBC (1995) which show that in many instances the outer streams such as Monoman Creek receive much larger salt loads than the inner ones such as Chowilla Creek (as they are the first to intercept the incoming regional groundwater). Furthermore, the salt loads predicted from the 1990/1991 flood are not that much greater than those from a flood almost half the magnitude. Finally, the predicted salt loads at the end of both floods return to a magnitude similar to that thought to represent "steady-state" (non-flood) loads (Table 3).

4.2.4. Localised recharge scenario modelling

All of the above simulations assume that no diffuse or localised recharge occurred during flooding and that the only recharge to the aquifer occurred via the bank storage. Jolly et al. (1994) found that diffuse recharge over the floodplain was minimal due to the highly dispersive nature of the sodic surface clay (Coonambidgal Formation) which restricts infiltration of the low salinity floodwater. These authors proposed that the long (18 months) salt recessions are due to the existence of isolated areas on the floodplain where the Coonambidgal Formation is either very thin or absent. When these areas are flooded, localised recharge occurs, resulting in the formation of a groundwater mound. As the mound dissipates in the months following flooding, saline groundwater stored in the zone between the location of the mound and the nearest creek is displaced into the creek.

To test this hypothesis, several simulation runs with a 100 000 ML day^{-1} peak flow were carried out with localised recharge occurring over a transect length of 692 m at a site centred 828 m to the north of Monoman Creek. Shown in Fig. 12 is a comparison of the predicted salt loads to Monoman Creek during the flood with no

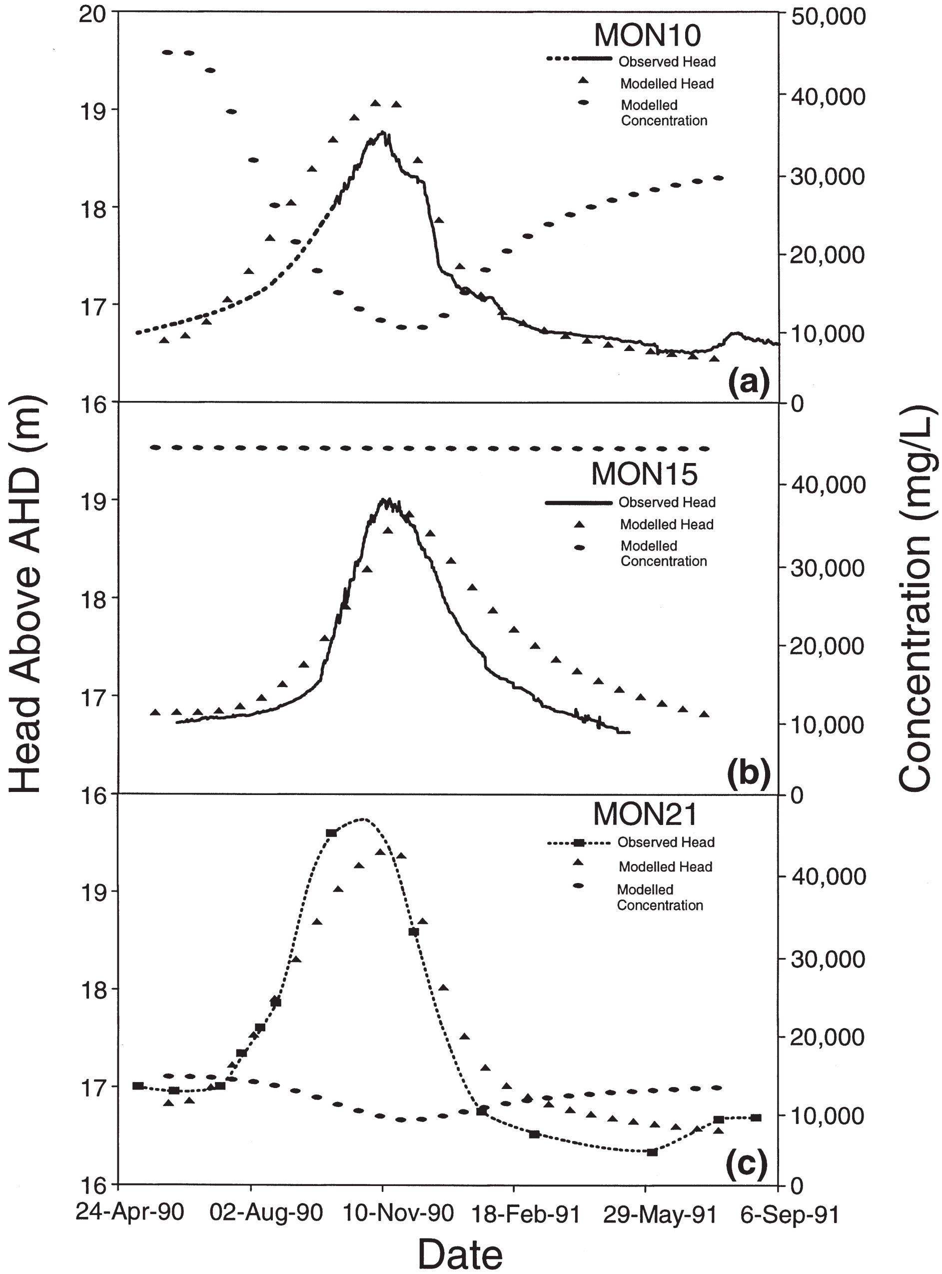

Fig. 8. Modelled and observed equivalent freshwater heads (in metres above the Australian Height Datum) and modelled concentrations during the 1990/1991 flood at piezometers (a) close to Monoman Creek (MON10), (b) in the centre of Monoman Island (MON15), and (c) close to Chowilla Creek (MON21).

localised recharge, with those with two different rates of localised recharge (2.0 and 5.0 mm day^{-1} for 80 days). These simulations illustrate that the length and magnitude of the salt load recession is extended by the presence of localised recharge, as hypothesised.

The predictions shown in Figs 11 and 12 were scaled over the entire lengths of Monoman (9.5 km) and Chowilla (11 km) Creeks to provide total salt output from this region of the Chowilla anabranch system. This was carried out on the model predictions for a 100 000 ML day^{-1} flood with no localised recharge, and for localised recharge at rates of 2 mm day^{-1} and 5 mm day^{-1} for 80 days. These are shown in Fig. 13, along with a smoothed curve of the salt load recessions for the 1981/1982 flood (peak flow 96 300 ML day^{-1}; Fig. 4). While an exact match between the observed and the modelled data is not expected (the observed data are for the whole of the Chowilla anabranch, whereas the modelled data are only for the Monoman and Chowilla Creek region), it can be seen that the inclusion of a component of localised recharge significantly enhances the degree to which the shape of the modelled salt loads match those of the observed salt recession.

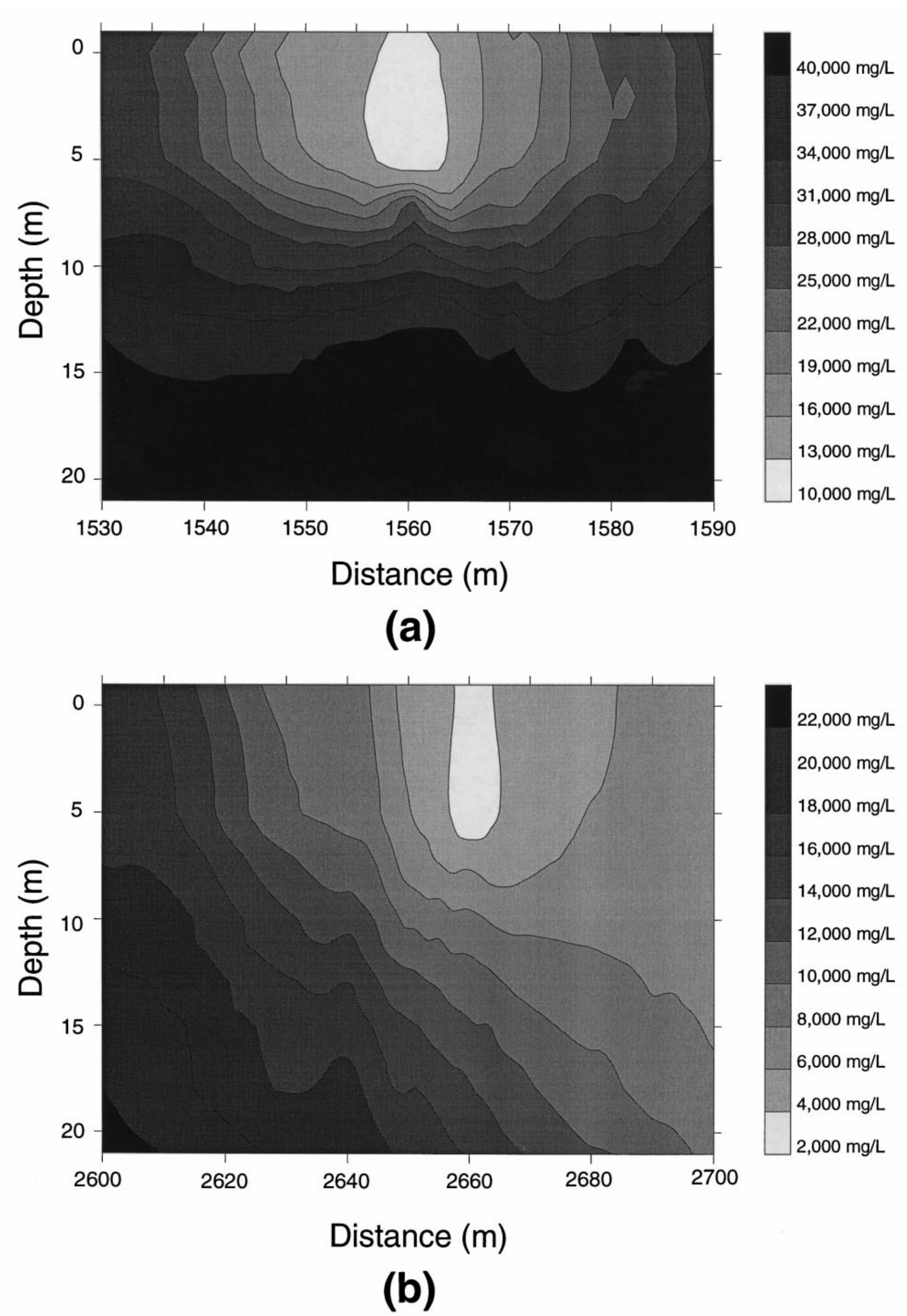

Fig. 9. Modelled concentration contours (mg L^{-1} TDS) in the vicinity of (a) Monoman and (b) Chowilla Creeks during the rising limb of the 1990/1991 flood (time step 600, t = 120 days).

5. Discussion and conclusions

It was not our intention to exactly replicate the salt loads observed during and after floods (the paucity of the field data does not allow an accurate comparison) but rather to simulate the general behaviour which is observed in the field. SUTRA was selected for its ability to describe density-dependent flow, to handle time varying boundary conditions, to account for unsaturated/saturated flow behaviour, and to allow for localised and diffuse recharge resulting from overbank flow. All four are important in describing the interaction between rising and falling stream levels and the adjacent groundwater. To our knowledge this is one of the first attempts to apply a density-dependent flow and transport model with time varying boundary conditions to stream/aquifer interaction.

The ease with which time varying boundary con-

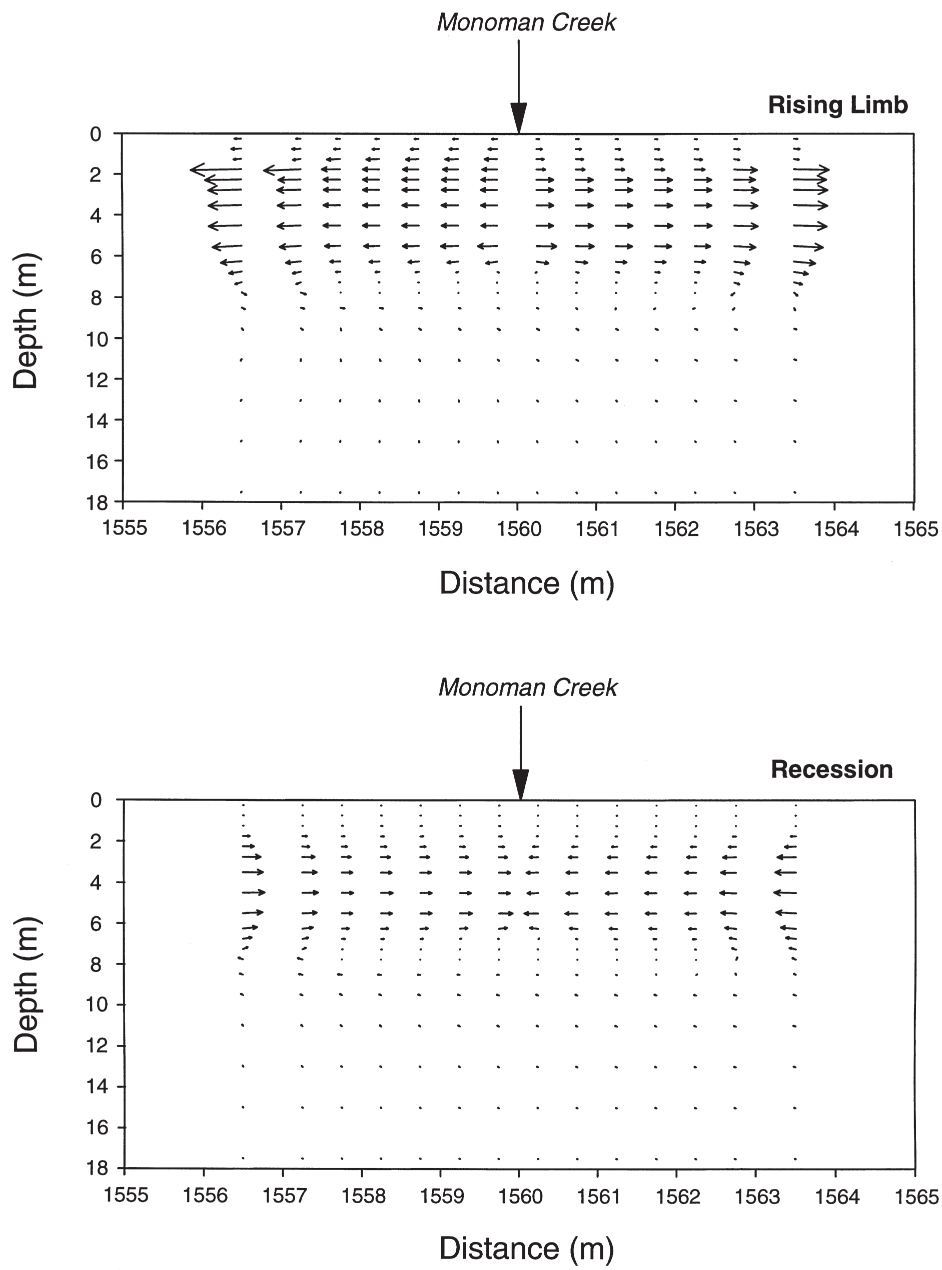

Fig. 10. Modelled velocity vectors in the vicinity of Monoman Creek during the rising limb (time step 600, t = 120 days) and recession (time step 1400, t = 280 days) of the 1990/1991 flood. The vectors are scaled linearly with maximum values of 9.0×10^{-6} m s^{-1} and 3.9×10^{-6} m s^{-1} for the rising limb and recession, respectively.

ditions can be defined (either in functional form and/or as discrete data) is one of the features which allows the use of SUTRA in studies of stream/aquifer interaction. While modifications to the program code are required, these are straightforward and well documented in Voss (1984). We were able to successfully model the time varying characteristics of two different stream boundaries simultaneously (Fig. 6).

Parameterisation of the hydraulic behaviour of the clay linings around each of the creeks (in particular Monoman Creek) was more problematic. In our initial simulation attempts we carried out trial and error manipulation of the specific pressure storativity of the clay lining (by varying the aquifer compressibility) to simulate the changes in degree of confinement of the aquifer in this region which occur as the groundwater heads rise and fall in response to the varying stream levels during a flood. Despite numerous calibration attempts there was generally poor matching (both in shape and timing) of the predicted and measured heads at the piezometers along the model transect. Even when reasonably good hydraulic matches were obtained, the

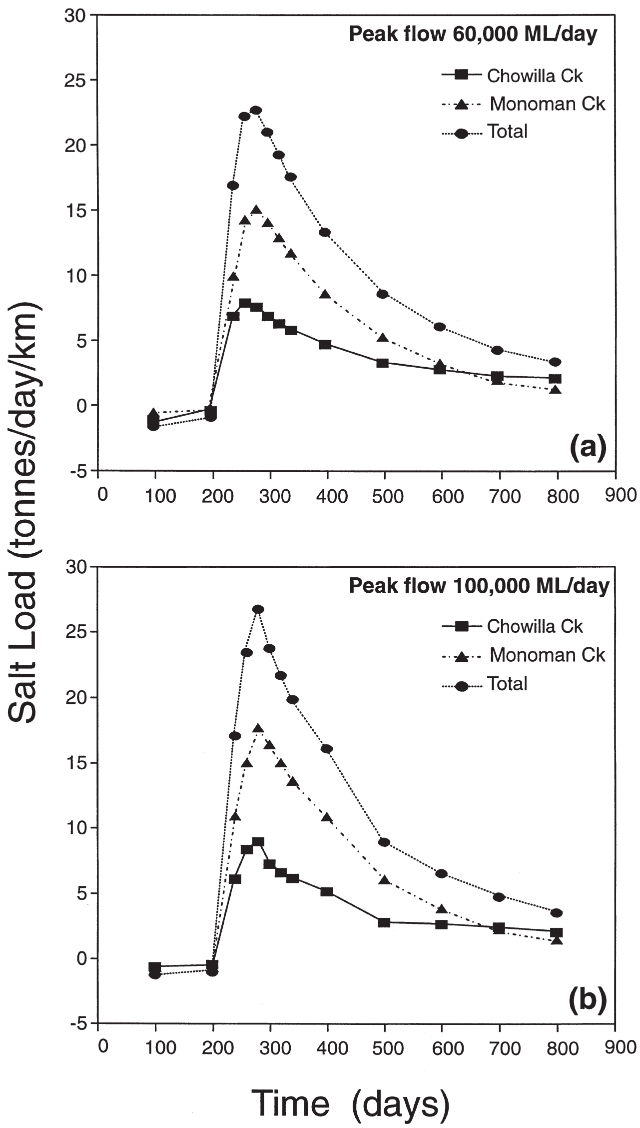

Fig. 11. Modelled salt loads to Monoman and Chowilla Creeks during (a) 60,000 ML day^{-1} and (b) 100,000 ML day^{-1} floods assuming no localised recharge.

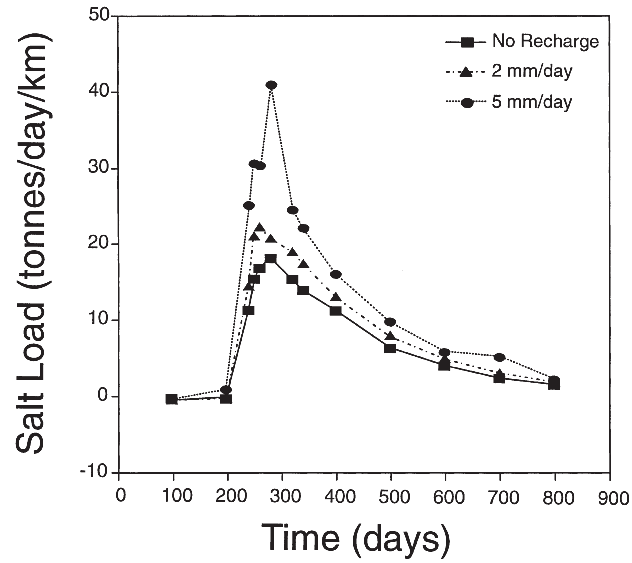

Fig. 12. Comparison of the modelled salt loads to Monoman Creek during a 100,000 ML day^{-1} flood with no localised recharge and two rates of localised recharge occurring over a transect length of 692 m centred 828 m north of Monoman Creek for 80 days.

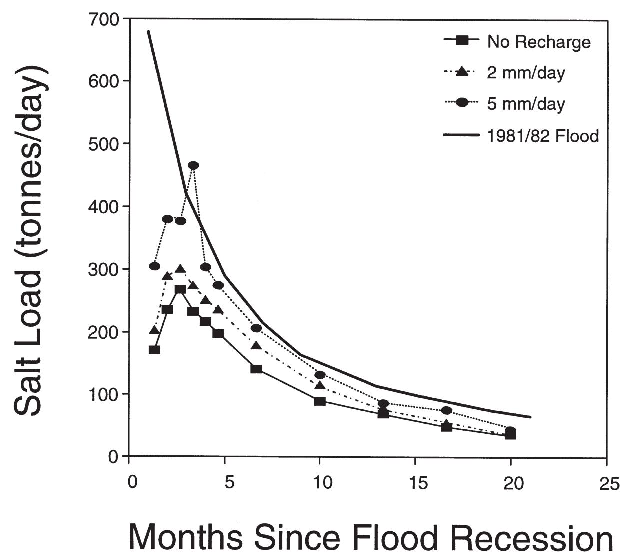

Fig. 13. Modelled salt load recessions (Monoman Creek plus Chowilla Creek) for a 100,000 ML day^{-1} flood with no localised recharge and with two rates of localised recharge over a transect length of 692 m centred 828 m north of Monoman Creek for 80 days. Also shown is the salt load recession for the whole of the Chowilla anabranch for the 1981/1982 flood which had a peak flow of 96,300 ML day^{-1} (as given in Fig. 4).

variations in the salinities in the aquifer adjacent to the streams were rarely close to those indicated by the field measurements. It became clear that it was necessary to have a more realistic representation of the saturation and de-watering of the unsaturated zone in the vicinity of the stream–aquifer boundaries. It was decided that an attempt would be made to utilise the saturated–unsaturated facility in SUTRA, and the mesh was modified to provide the more detailed discretisation required for numerical stability of the unsaturated flow and solute transport. Values appropriate for a silty clay loam soil (based on previous work at the study site by Jolly et al., 1993) were used for the three Van Genuchten (1980) parameters. Considerably better matches between the predicted and measured heads and concentrations were then obtained.

Once successful parameterisation of the clay lining around the creeks was achieved, the model provided good predictions of the piezometric response of the groundwater to the rising and falling stream levels (Fig. 8). While we did not have continuous measurements of the groundwater salinity near the creeks, the predicted freshening of groundwater around each of the creeks during the flood (Figs 8 and 9) is consistent with the limited field data (see Jolly et al., 1994). Moreover, the predicted velocity vectors during the flood recession indicate that the greatest fluxes of water, and hence salt, returning to the creeks from bank storage occur near the base of the creeks, as suggested by the field data presented in Fig. 7.

The predictions in Fig. 11 suggest that Monoman Creek receives greater salt loads following floods than does Chowilla Creek. This is consistent with the observations reported in MDBC (1995) which show that the outer streams such as Monoman and Punkah Creeks (see Fig. 1) act as "interception drains" for the saline groundwater flowing into the floodplain from the north from the regional unconfined Pliocene Sands aquifer. These predictions also show that bank storage mixing only accounts for the 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, both in terms of duration (Fig. 12) and shape of the salt recession curve (Fig. 13). These results appear to corroborate the hypothesis of Jolly et al. (1994) that localised recharge on the Chowilla floodplain during large floods is responsible for the long salt recessions in the floodplain streams.

SUTRA appears to be an effective tool for modelling stream–aquifer interaction in situations where there is a large salinity contrast between the stream and the aquifer. In the field situation presented here it provided a good representation of both water and solute movement to/from and within the aquifer during and after floods. Based on the success of this application of SUTRA we conclude that the methodology can be used to investigate similar cases.

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