# The Impact of Flooding on Modelling Salt Transport Processes to Streams

Ian D. Jolly[1], Kumar A. Narayan[1], Don Armstrong[2] and Glen R. Walker[3]

[1]CSIRO Land and Water, PMB 2, Glen Osmond, SA, 5064
[2]Lisdon and Associates, 77 Hawthorndene Drive, Glenalta, SA, 5052
[3]CSIRO Land and Water, GPO Box 1666, Canberra, ACT, 2601

**Abstract** The movement 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 2-D, variable density, finite element flow and solute transport model (SUTRA). The model was applied in cross-section over a 6 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 was required, as hypothesised by previous field-based studies.

## 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 (eg. 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 stream salinisation. This extensive river system drains an area ~$10^6$ km$^2$ (one-seventh the area of the Australian continent), most of which 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 to 800 $\mu$S cm$^{-1}$. The large increases in this reach are due to the discharge of saline (>20,000 $\mu$S cm$^{-1}$) groundwater from regional aquifers to the river, an essentially natural process exacerbated by irrigation activities in close proximity to the river (Evans, 1988). Of considerable concern is that the salinity of the River Murray in this reach is increasing at a rate >2% per year since 1970 (Morton and Cunningham, 1985).

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. 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).

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 (eg. Hall and Moench, 1972) and numerical (eg. Marino, 1981), have been developed for the study of stream-aquifer interaction. While many of the models account 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. Morover, a phenomenon often typical of stream-aquifer interaction in arid and semi-arid areas is a large salinity contrast 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 (eg. Ghassemi et al., 1989). However, studies such as those of Herbert et al. (1988) suggest that large salinity contrasts can result in density-dependent flow, a phenomenon rarely considered in studies of stream-aquifer interaction.

In this paper we use a 2-D variable density flow and solute transport model (SUTRA; Voss, 1984) to better understand the processes operating in the transport of saline groundwater to floodplain streams. Besides density effects, the SUTRA model can be modified to handle time varying boundary conditions typical of stream-aquifer behaviour. 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 from the Chowilla floodplain in South Australia. 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. STUDY AREA

The Chowilla anabranch, a 200 km$^2$ semi-arid floodplain situated near the junction of the South Australian, Victorian and New South Wales borders was the focus of this study; in particular the region in the vicinity of Monoman Island and Lock 6 (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. Prior to the commencement of flow regulation in the 1920-30's, 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. Under present-day conditions Lock 6 creates a permanent upstream weir pool which drives water continually through the Chowilla floodplain streams which act as interception drains for the saline regional groundwater system (Fig. 1). Moreover, inundation of the floodplain causes large volumes of saline groundwater to flow into the creeks following a major flood with elevated salt loads often continuing for up to two years after a flood event (NEC, 1988), with the resulting increases in River Murray salinities having detrimental effects on downstream water users.

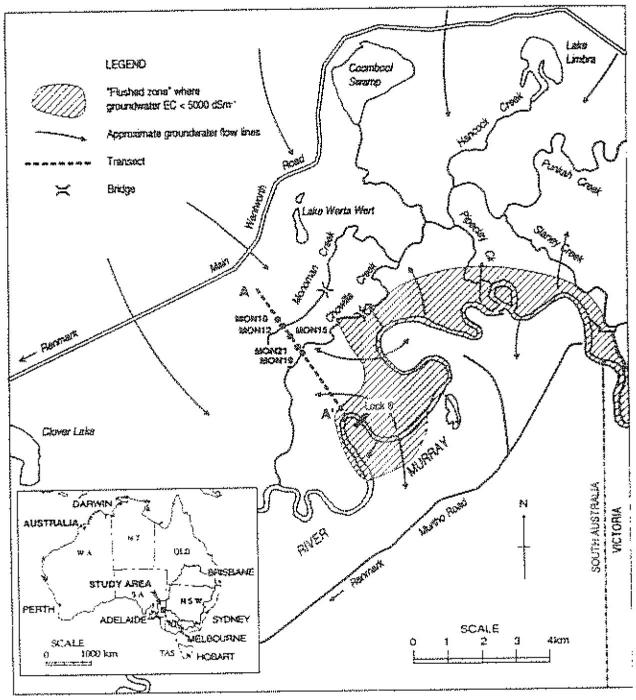

Figure 1: The Chowilla anabranch study area.

A typical hydrogeological cross-section of the floodplain is shown in Fig. 2. The main aquifers are the Monoman Formation and the regional Pliocene Sands aquifer. The Monoman Formation consists of alluvial sands approximately 30 m in thickness with hydraulic conductivities in the range 4-10 m day$^{-1}$. The water table formerly resided in this aquifer, but due to the construction of Lock 6 it has risen 2-3 m and is now often found within the overlying Coonambidgal Clay. Salinities in the aquifer range from 20,000 to 60,000 mg L$^{-1}$ with the highest values at the base (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 (<3,000 mg L$^{-1}$) 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 Pliocene Sands aquifer consists of estuarine fine, medium and coarse sands with some clayey layers at the base of the aquifer. The aquifer is unconfined and has salinities in the range 40,000-90,000 mg L$^{-1}$. The hydraulic conductivity is lower than that of the Monoman Formation and is of the order of ~5 m day$^{-1}$. Regional groundwater flow occurs laterally from the Pliocene Sands aquifer towards the floodplain, entering the Monoman Formation from the north.

The soils of the Chowilla floodplain generally consist of a micaceous cracking clay of alluvial origin (Coonambidgal Clay) which 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 Clay 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 Clay overlies the Monoman Formation.

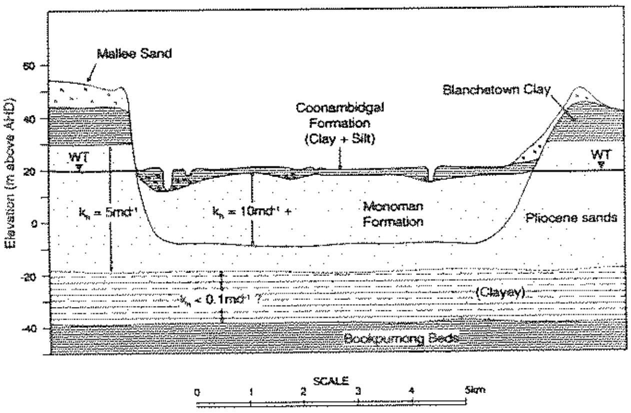

Figure 2: Typical hydrogeological cross-section.

## 3. MODEL DESCRIPTION

The SUTRA (Saturated-Unsaturated Transport) model was used to numerically simulate changes in the aquifer pressure and salt concentration in response to flooding along a vertical slice through the floodplain (transect A-A'; Fig. 1). Several piezometers which lie along the transect were used in the model calibration. The model employed a 2-D finite element approximation of the governing equations in space and an implicit finite difference approximation in time. For solute transport, two partial equations were solved simultaneously (fluid and salt mass-balance equations) to compute the pressure and solute concentration at each node for each time step. The governing equations of flow and solute transport are given in Voss (1984). In order to compare the predictions of pressure with field measurements of piezometric head, the pressures were transformed to "equivalent freshwater heads" ($h = z + p/\rho g$).

To represent the major features of the system the mesh was discretised in a non-uniform quadrilateral manner with 2914 nodes and 2790 elements (Fig. 3). A fine mesh spacing of 0.5 m adjacent to 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 head boundary conditions for the simulation are also shown in Fig. 3. Inflow occurring through the northern specified head boundary had a concentration of 45,000 mg $L^{-1}$ TDS and from the river end of 350 mg $L^{-1}$ TDS. Any flow out of the mesh at the specified head boundaries occurred at the ambient concentration of the aquifer fluid. Solute neither dispersed nor advected across no-flow boundaries. Values used for input into the SUTRA model are given in Table 1.

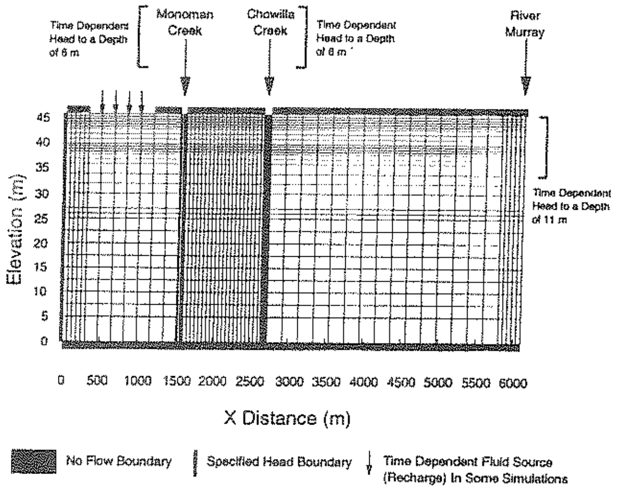

Figure 3: Model mesh and boundary conditions.

In the case of the time varying boundary conditions, the River Murray 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. Sinusoidal functional approximations of the time-dependent boundary conditions during flooding were incorporated in the BCTIME subroutine of SUTRA and produced good matches with the observed hydrographs (Fig. 4).

The saturated/unsaturated mode of SUTRA was applied using the three parameter equations of van Genuchten (1980) which relate pressure to percentage saturation and unsaturated hydraulic conductivity. Values typical of a silty clay loam were used for the unsaturated zone parameters (Table 1; note that $S_{WRES}$, $A_A$ and $V_N$ in the SUTRA model correspond to $\theta_r$, $\alpha$ and n in van Genuchten, 1980). 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.

**Table 1: Model parameters.**

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Freshwater density ($\rho$) = 1000 kg $m^{-3}$
Groundwater density ($\rho_{gw}$) = 1030 kg $m^{-3}$
Fluid viscosity ($\mu$) = $10^{-3}$ kg $ms^{-1}$
Coefficient of fluid density change ($\delta\rho/\delta C$) = 700 kg $m^{-3}$
Water compressibility ($\beta$) = $4.5\times10^{-10}$ $Pa^{-1}$
Porosity ($\varepsilon$) = 0.30
Hydraulic conductivity of Upper Monoman Sands ($K_h$) = 10 m $day^{-1}$
Hydraulic conductivity of Lower Monoman Sands ($K_h$) = 4 m $day^{-1}$
Hydraulic conductivity of Loxton/Parilla Sands ($K_h$) = 4 m $day^{-1}$
Ratio of vertical to horizontal hydraulic conductivity ($K_v/K_h$) = 0.1
Longitudinal dispersivity ($\alpha_L$) = 20 m
Transverse dispersivity ($\alpha_T$) = 1 m
Molecular diffusivity ($D_m$) = $1.5\times10^{-9}$ $m^2 s^{-1}$
Residual saturation ($S_{WRES}$) = 0.1
Van Genuchten (1980) parameter ($A_A$) = $5\times10^{-5}$ m $s^{-2}$ $kg^{-1}$
Van Genuchten (1980) parameter ($V_N$) = 2.0
Upper Monoman Sands thickness = 6 m
Lower Monoman Sands thickness = 20 m
River Murray depth = 11 m
Monoman Creek depth = 6 m
Chowilla Creek depth = 6 m

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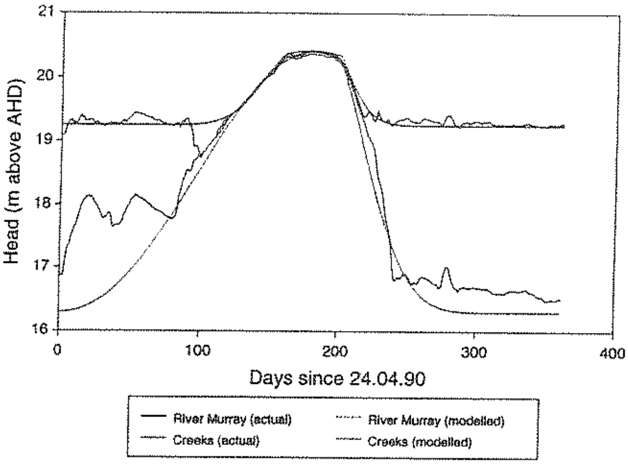

Figure 4: Time varying boundary conditions for a River Murray flood with a peak flow of 100,000 ML $day^{-1}$.

## 4. SCOPE OF THE SIMULATIONS

In order to produce realistic initial pressure and concentration conditions for later 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 $L^{-1}$, and a constant head boundary of 19 m AHD at the northern end at a groundwater salinity of 45,000 mg $L^{-1}$. 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.

For the flooding scenarios, calibration of the model was carried out in a number of steps. This was necessary because the nature 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 during a 100,000 ML day$^{-1}$ flood in 1990/91. Furthermore, a sensitivity analysis of the aquifer storage coefficient around Monoman and Chowilla Creeks was carried out. 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. All flooding scenarios were run for 4000 time steps, each of 0.2 of a day duration, giving a total simulation time of 800 days.

## 5. RESULTS

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, suggesting that the storage coefficients around the creeks differ as well. A number of simulations of the 1990/91 flood were carried out with the hydraulic conductivity of the Chowilla Creek clay lining fixed at $6x10^{-2}$ m day$^{-1}$ and that around Monoman Creek varying between $6x10^{-4}$ and $1.2x10^{-2}$ m day$^{-1}$. Comparison of the model predictions with measured data were made at three piezometers along the transect (one close to Monoman Creek, **MON10**; one in the centre of Monoman Island, **MON15**; and one close to Chowilla Creek, **MON21**). All three piezometers provided reasonable matches to the magnitude and timing of the peak of the hydrograph (Fig. 5). However, the shape of the hydrograph is not modelled as well, although the slightly asymmetric rise and fall is simulated. From these and the results for other piezometers (data not shown here), a hydraulic conductivity of $6x10^{-4}$ m day$^{-1}$ was chosen as being appropriate for the clay lining around Monoman Creek.

Using the parameters determined above the model was run to simulate the 1990/91 flood assuming no diffuse recharge. Shown in Fig. 5 are the predicted and observed heads and the predicted concentrations at nodes representing piezometers MON10, 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.

Shown in Fig. 6 are the predicted salt loads entering Monoman and Chowilla Creeks during the course of the 1990/91 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 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 1990/91 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.

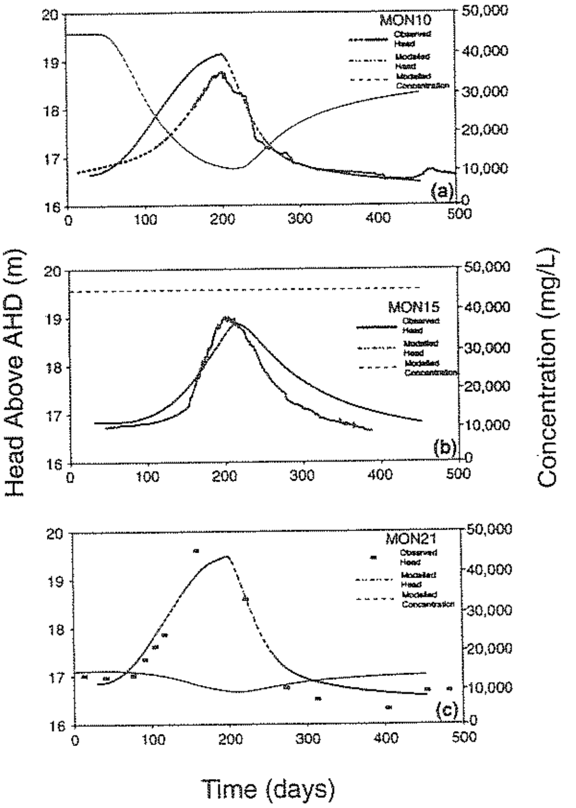

**Figure 5: Modelled and observed heads and modelled concentrations at three piezometers during the 1990/91 flood.**

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 Clay) 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 Clay 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.

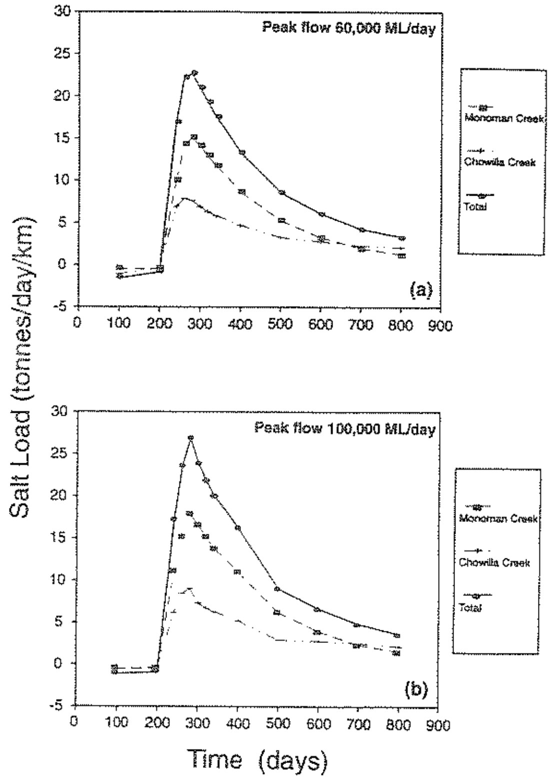

**Figure 6: Modelled salt loads to Monoman and Chowilla Creeks during a 60,000 and 100,000 ML day$^{-1}$ flood.**

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. 7 is a comparison of the predicted salt loads to Monoman and Chowilla Creeks (scaled over their entire lengths) during the flood with no localised recharge, with those with two different rates of localised recharge (2 and 5 mm day$^{-1}$ for 80 days). Also shown is the actual salt load recession for a similar size flood in 1981/82 (peak flow 96,300 ML day$^{-1}$). While an exact match between the observed and the modelled data is not expected (the observed data is for the whole of the Chowilla floodplain, 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. These simulations illustrate that the length and magnitude of the salt load recession is extended by the presence of localised recharge, as hypothesised in the field study of Jolly et al. (1994).

## 6. DISCUSION AND CONCLUSIONS

It was not our intention to exactly replicate the salt loads observed during and after floods (the paucity of the field data did not allow an accurate comparison) but rather to simulate the general behaviour which was 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 was one of the first attempts to apply a density-dependent flow and transport model with time varying boundary conditions to stream/aquifer interaction, on a saline floodplain.

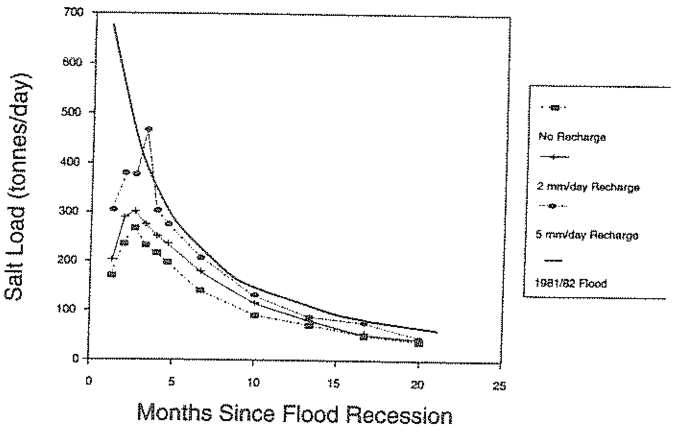

**Figure 7: Modelled salt load recessions (Monoman Creek plus Chowilla Creek) for a 100,000 ML day$^{-1}$ flood with no localised recharge and with two rates of localised recharge. Also shown is the salt load recession for the whole of the Chowilla anabranch for a similar size flood in 1981/82.**

The ease by which time varying boundary conditions 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. 4).

Modelling of the saturation and de-watering of the unsaturated zone in the vicinity of the stream boundaries 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 variations in the near-stream salinities in the aquifer were rarely close to those indicated by the field measurements. It was then 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 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. 5). 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 was consistent with the limited field data (see Jolly et al., 1994). Moreover, the predicted velocity vectors (not shown here) 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 field data (not shown here).

The predictions in Fig. 6 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 regional unconfined Pliocene Sands aquifer to the north. 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 and shape of the salt recession curve (Fig. 7).

In conclusion, 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. In particular, the model outputs 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.

## 7. ACKNOWLEDGEMENTS

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