Solute and Water Movement in Irrigation Bays: A Predictive Management Model

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Abstract Salinisation of Australia's inland waterways is a significant problem and its management involves significant financial resources. A key component in understanding catchment salt export is interpreting the on-land processes which determine runoff and groundwater recharge/discharge. This paper presents a predictive model for water and salt movement within irrigation bays. The bay model employs a physically-based formulation, where: groundwater movement is explained by Darcy's equation, capillary rise by an approximate solution to the Kirchoff transformed Richards equation, and overland flow by a Mannings type equation. These process descriptions are solved using the Laplace transform/modified finite analytic method, a quasi-analytical method but with the flexibility of more traditional numerical methods. The bay model is tested through application to a field site where water and salt transport processes were intensively monitored over a two year period. Model predictions of overland flow and transport, groundwater and unsaturated water movement are shown to be in close agreement with field observations. The bay model is part of a catchment salt export management model currently under development.

1. INTRODUCTION

In the irrigation areas of northern Victoria and southern New South Wales irrigation has lead to saline watertables close to the soil surface. Not only does this saline groundwater pose a risk to plant productivity, but significant volumes of salt are exported into the river system. Management of these problems is aimed at achieving a sustainable balance between the on-farm and river water quality objectives.

The predominant irrigation method in the irrigation areas of northern Victoria and southern New South Wales is flood, or border, irrigation. Large areas have been laid out into irrigation bays and surface drainage systems established. Laser grading, where the bay surface is formed to a specific slope, has been used extensively. In a surface water sense the bays are hydrologically separate with flow divides at the top and bottom, and checkbanks down the sides. In the sub-surface, the shallow groundwater behaviour will primarily be driven by the surface water balance within the bay with linkages to deeper groundwater flow systems being a secondary but still important process.

A key part of determining the best management strategy for a given area is predicting the consequences of various management options. Predictive modelling of the water and salt movement processes allows the relative benefits of different strategies to be evaluated. Such modelling has to be able to relate the impact of a wide range of possible effects; from changes to irrigation practice to drainage. While models have been developed for irrigation front advance and therefore include a representation of infiltration behaviour, little work has been done on explaining the links between overland flow and the groundwater system and the resultant salt transport.

The SHE model (Abott et al, 1986) has been applied to explain the salt and water movement processes in irrigation bays in the Tragowel plains area of northern Victoria (Mudgway et al, 1997; Jayatilaka et al, 1997). An important aspect of the soils at the site was their cracking nature, which invalidated the porous media assumptions employed in SHE. More importantly SHE is a large, generalised, catchment model and as such would be difficult to apply in a routine management fashion. To be tractable, a management model needs to be taylored to the problem of interest, thus avoiding unnecessary complexities and ensuring the process descriptions are appropriate. In addition, such a model must be compatible with end user computing resources, available data, and technical capabilities.

This paper describes a physically based water flow and salt transport model for irrigation bays. The objective of the model is to provide a framework for investigating questions of irrigation practice and bay design for different site characteristics. The model is tested through simulation of a field experiment in northern Victoria, Australia.

2. MODEL DERIVATION

2.1 Overview

Figure 1 presents the conceptualisation upon which the model is based. Previous work has shown that groundwater seepage and exfiltration are very important contributors to the surface water salt load and that groundwater had a dynamic behaviour (Mudgway and Nathan, 1993). Groundwater can rise rapidly during infiltration events and then recede in response to evaporation and groundwater movement. The unsaturated zone processes of infiltration and capillary rise promote groundwater movement and determine the amount of salt in the near surface zone, which can be readily available to surface waters. Wash-off of surface precipitated salt, the result of bare surface evaporation, is also a significant source of salt for surface waters.

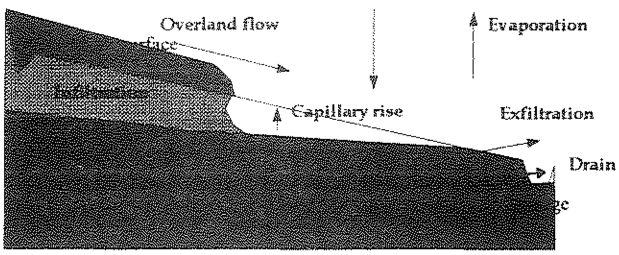

Figure 1: Irrigation bay conceptual model.

The flow processes can be divided into the following groups:

- Unsaturated zone (infiltration and capillary rise).
- Overland flow, and
- Groundwater movement.

Salt transport can occur in each of the above flow processes.

Another important criteria for the model is the computational burden and thus speed with which calculations can be performed. To maximise the utility of the model computational overheads had to be minimised. This was achieved through exploiting aspects of the flow process to simplify the process descriptions, particularly with the unsaturated zone, usually the most computationally demanding to solve.

The organisation of the bay model is presented in Figure 2. The bay is divided into a series of vertical columns in which vertical saturated/unsaturated flow is modelled. The width of the vertical columns is coincident with the spatial discretisations used for overland flow and groundwater. The watertable intersects the vertical saturated/unsaturated columns.

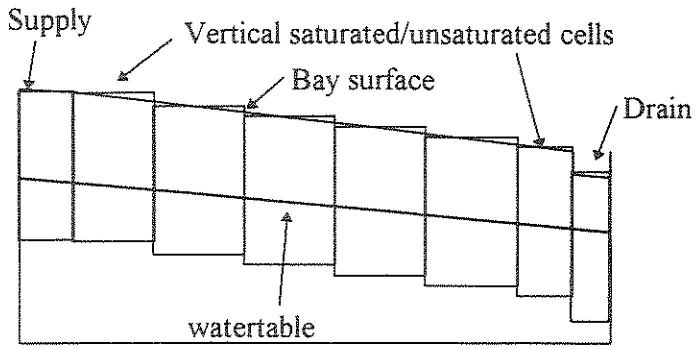

Figure 2: Bay model spatial discretisation.

2.2 Unsaturated flow

Saturated/unsaturated flow is treated separately for infiltration and capillary rise. It is assumed that capillary rise is dominated by soil matrix flow processes, whereas surface ponded infiltration may be a mixture of matrix and macropore flow processes.

2.2.1 Capillary Rise

In a shallow watertable environment, large recharge events such as irrigation, tend to saturate the soil profile and raise the groundwater table to the soil surface (Gilfedder et al, 1997). Capillary rise during watertable recession after a saturating event is conceptualised as being composed of two stages; at early times water movement within the soil is not limiting and the matric potential is directly proportional to depth. The duration of this stage depends on the soil hydraulic conductivity and the evaporation rate, where the more permeable the soil the longer this initial stage. The second stage is where water movement is determined by the soils ability to conduct water. Since capillary rise is a matrix flow process, an approach based on Richard's equation (de Marsily, 1987) is appropriate. Separating capillary rise into two stages reduces the computational load signficantly.

Equilibrium model of capillary rise

At equilibrium the matric potential profile is determined more by gravitational forces than water movement; when the rates of water movement are not limiting the evaporation rate. This assumption is exact when evaporation is zero where $\partial h/\partial z = 1$, from Darcy's law. However it is often the case in an evaporating soil that $\partial h/\partial z$ is close to unity, particularly for more premeable soils. When the soil conductivity is large in comparison to the evaporation rate, slight differences are enough to support water movement to the plant root zone at the rate at which it is removed.

The equilibirum assumption, $h=z-h_{surf}$ (where h is the matric potential, z depth, and h_{surf}, the matric potential at surface) is applied to the following mass balance relation,

$$R\,\Delta t = \int_0^{\ell} \theta dz \Big|_{t+\Delta t} - \int_0^{\ell} \theta dz \Big|_{t} \quad (1)$$

where R is the evaporative loss from the unsaturated zone, θ is the volumetric water content, ℓ is some maximal depth into the saturated zone and is constant, and Δt the time step.

Equation (1) can also be written as,

$$R\,\Delta t = \theta_s(h_{surf1} - h_{surf2}) + \int_{h_{surf1}}^{h_{surf2}} \theta dh \quad (2)$$

where h_{surf1} is the matric potential at the soil surface at time t, h_{surf2} is the potential after the timestep, and θ_s is the saturated water content. Since $h = (z-h_{surf})$, h_{surf} is equivalent to the negative of the depth to groundwater. Given R, h_{surf1} can be determined at each time step through a numerical root finding procedure.

Figure 3 presents a comparison between the equilibrium model results and those obtained from a finite element solution (using 100 nodes over a 100cm column) of the saturated/unsaturated Richards/Darcy equation for Guelph loam properties (where K_s, the saturated hydraulic conductivity, is equal to 1.32 cm/hr). Figure 4 presents results for Grenoble sand (K_s = 15.32 cm/hr). The values of the physical properties for Guelph loam and Grenoble sand properties can be found in Connell (1997).

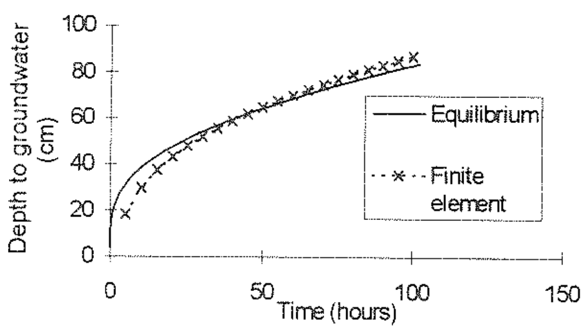

Figure 3: Depth to watertable vs time for equilibrum model and finite element solution procedures for Guelph loam with an evaporation rate of 8 mm/day and a root depth of 20 cm.

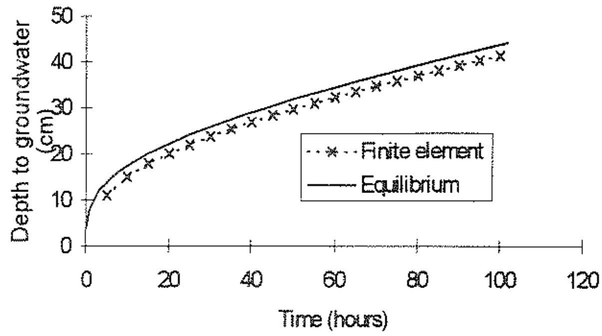

Figure 4: Watertable recession for Grenoble sand with an evaporation rate of 8mm/day and a root depth of 20 cm.

Approximate Richards Equation Solution

For soils with lower permeability, and at longer times during recession for more permeable soils, an approximate solution to Richards equation was developed with the root zone treated in a mass balance fashion. A full description of the capillary rise solution procedure can be found in Connell (1997). The approach is based on the quasilinear form of Richards' equation for the unsaturated zone and Darcy's equation for the saturated zone. The generalised, transformed flow equation can be written as,

$$\frac{\partial \phi}{\partial t} = \gamma\left[\frac{\partial^2 \phi}{\partial z^2} - \mu\frac{\partial \phi}{\partial z} - R\right] \quad (3)$$

where for saturated flow, $\phi = K_s(h'+1/\alpha)$, $\gamma = K_s/S_f$ and $\mu = 0$. For unsaturated flow, $\phi = \int_{-\infty}^{h} K(\bar{h})\,d\bar{h}$, $\gamma = D$ and $\mu = \alpha$. Where S_f is the specific storage coefficient, D is the linearised soil moisture diffusivity, and h' is the hydrostatic pressure. The unsaturated hydraulic conductivity is defined as $K(h) = K_s \mathrm{Exp}(\alpha h)$ where α is the hydraulic conductivity shape parameter.

This equation is solved using the Laplace transform/modified finite analytic method (described in Connell et al, 1997). The key assumption with this technique is that the diffusivity, D, in the unsaturated zone is constant within a timestep. The diffusivity is a key property which relates the release/storage of water from the soil pore space as the soil desaturates or resaturates as well as the rate of water movement. The approach employed is to update the diffusivity at the end of each timestep. The diffusivity is therefore approximated in a piecewise continuous fashion.

The current soil hydraulic properties implemented are the van Genuchten moisture retention relationship (van Genuchten, 1980),

$$\theta = \theta_r + \frac{\theta_s - \theta_r}{\left[1 - \left(\frac{h_g}{h}\right)^n\right]^m}$$

where θ_r is the residual water content, h_g is the van Genuchten scale parameter, and n,m are shape parameters.

Water content variation in the root zone is described by the following relation.

$$\frac{\partial \theta}{\partial t} = R - q_{CR} / z_{roots} \quad (4)$$

where q_{cr} is the root zone flux calculated by solving (3), and z_{roots} is the rooting depth.

Equation (4) is solved in an explicit time stepping manner. The value of the root zone water content is used as the boundary condition in for solution of equation (3).

2.2.2 Infiltration

In soils where macropores are present infiltration can have a large initial step as these fill up. This 'crack fill' amount will either go directly to the groundwater system, bypassing the unsaturated zone, or move laterally into the soil matrix from the macropore. After the sharp initial step in infiltration rate, the rate is observed to be relatively constant. This behaviour is currently implemented in the model, however any infiltration function can be incorporated. Infiltrated water is added to the unsaturated profile. In the current implementation, the unsaturated zone is 'filled up' from the watertable upwards. Once the profile is saturated, infiltration is determined by groundwater movement.

2.3 Unsaturated Solute Transport

Solute transport within the unsaturated/saturated vertical columns is described by a form of the advection-dispersion equation. For unsaturated porous media this can be written as,

$$\frac{\partial(\theta C)}{\partial t} = \frac{\partial}{\partial z}\left[\theta D_m \frac{\partial C}{\partial z}\right] - \frac{\partial(qC)}{\partial z} + R_{sol} \quad (5)$$

where C is the concentration of solute, D_m is the dispersion coefficient, and R_{sol} is a sink-source term for solute mass. Expanding the derivative terms in (5) and introducing the continuity equation for water leads to,

$$\theta \frac{\partial C}{\partial t} = \frac{\partial}{\partial z}\left[\theta D_m \frac{\partial C}{\partial z}\right] - q\frac{\partial C}{\partial z} + R_{sol} - CR \quad (6)$$

Equation (6) must be linearised, with constant coefficients, to apply the Laplace transform/modified finite analytic method. Equation (6) is therefore rewritten as,

$$\frac{\partial C}{\partial t} = D_m \frac{\partial^2 C}{\partial z^2} - \frac{q}{<\theta>}\frac{\partial C}{\partial z} + \frac{R_{sol}}{<\theta>} - C\frac{R}{<\theta>} \quad (7)$$

where $<\theta>$ is a constant, in the case of the Laplace transform/modified finite analytic method it is an average for a timestep and element. A similar equation to (7) was solved in Connell and Haverkamp (1996) using the Laplace transfrom/modified finite analytic method. This paper presents an analysis of the accuracy of the solution procedure as well as more details of the overall methodology.

2.4 Groundwater Flow

Groundwater movement in the model is described by Darcy's equation with the Dupuit approximation, which means that flow is assumed to be parallel and horizontal. While the bays have a small slope the aquifer base will most probably be horizontal for most model applications. For this situation the Dupuit equation is more representative of the flow process than the Boussinesq equation.

The confined Darcy-Dupuit equation can be written as,

$$S\frac{\partial h}{\partial t} = \frac{\partial}{\partial x}\left[K_s \sigma \frac{\partial h}{\partial x}\right] - R(x,t) \quad (8)$$

for one dimensional flow (de Marsily, 1986). In equation (1), h is the hydraulic head (L), K_s, the hydraulic conductivity (L/T), S the specific storage coefficient, σ is the aquifer thickness (L), R is the recharge/discharge/leakage (L/T), and x is the horizontal coordinate (L).

The groundwater solution scheme and its performance is described in Connell et al (1997).

2.5 Overland flow

Overland flow, Q, is described using a simple mass balance approach combined with a discharge depth relation of the following form (from Maheshwari and McMahon, 1992),

$$Q = d_1\, y^{d_2} \sqrt{S_e} \quad (9)$$

Where $S_e = S_w - S_o$, with S_o being the surface slope, S_w the slope of the water surface, y is the ponded depth and d1, d2 are physical properties. If $d_2 = 5/3$ equation (9) is Manning's equation.

The surface of the bay is divided into elements or cells, and within each cell a simple mass balance is applied. That is,

$$\frac{\partial y_i}{\partial t} = (Q_{i-1} - Q_i) / \Delta x_i - \frac{dI_i}{dt} - \frac{dE_i}{dt} + R_{ain} \qquad (10)$$

where I_i is the cumulative infiltration, E_i is cumulative evaporation, and R_{ain} is the rainfall rate.

In a similar manner to water movement, a mass balance relation for solute can be written as,

$$\frac{\partial (yC)}{\partial t} = -\frac{\partial}{\partial x}(QC) + R_c \qquad (11)$$

where C is the concentration of solute in overland water and R_c is the sink/source of solute.

3. APPLICATION

The model was tested through application to the field experiment described by Gilfedder et al (1997). Soil hydraulic properties were measured through Talsma tubes and infiltration tests using a large diameter infiltration ring. Properties for the overland flow relation, equation (10), were approximated with values from Maheswari and McMahon (1992).

The good agreement between observations and model results for overland flow (Figure 5) demonstrate that not only the overland flow properties are correctly estimated but that infiltration is also accurately represented. There are differences between the model and observations during the later surface water recession stages; a result of the highly heterogeneous flow path.

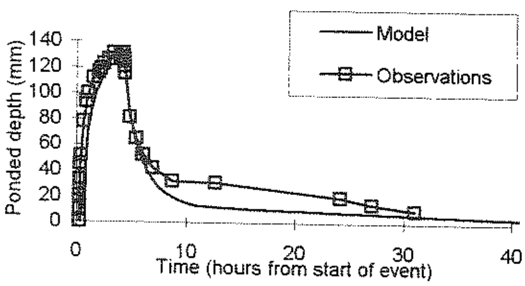

Figure 5: Model results and observations for ponded depth during an irrigation event 25m from the supply channel.

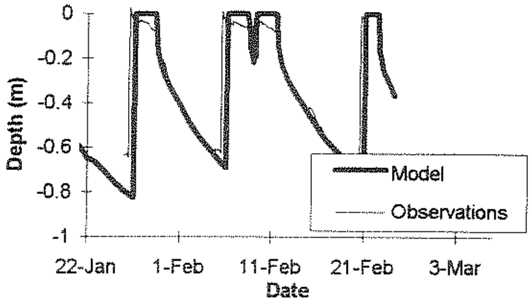

Figure 6: Model simulation of groundwater elevations compared with observations for 130m from supply channel.

The dynamic response of the groundwater system to irrigation events was correctly simulated by the model (see Figure 6). The rapid rise to the soil surface as the soil profile fills, and at the end of ponding the groundwater recession behaviour are accurately simulated.

In Figure 7 the model is shown to be able to accurately simulate field observations of drain flux with time. There are small differences in terms of estimating the time of arrival of the irrigation front, but the magnitude of the peak flow, and the recession behaviour are well represented.

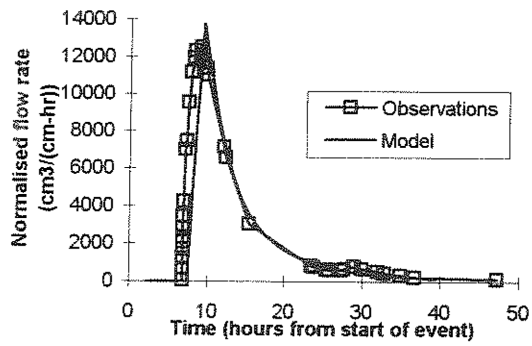

Figure 7: Simulation of drain flow with respect to time compared with field observations.

Figure 8 presents model results and observations for cumulative drain flow for three irrigation events. The first two events involve large supply volumes and fairly high inefficiencies. For the last event the irrigation supply volume was small and irrigation efficiency much higher. The good agreement between model results and observations for these three events indicates that the model is able to represent a range of flow and transport conditions.

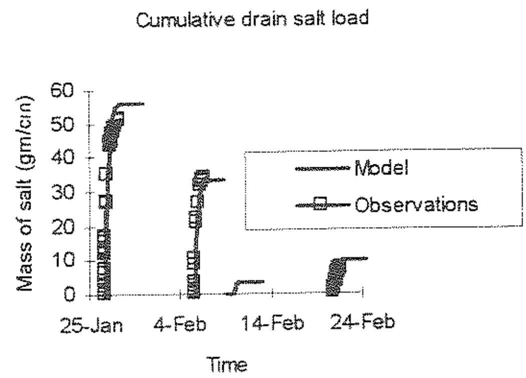

Figure 8: Model results and observations of cumulative drain salt mass for three consecutive irrigation events.

4. CONCLUSIONS

This paper has presented results from an investigation of the processes operating within irrigation bays which could lead to salt export. A predictive model was developed based on the observations from a field experiment in the Cohuna area of northern Victoria. This model was tested by simulating observations from the field experiment and was found to be in close agreement. The model can readily be used to predict the consequences of changes in irrigation practice, irrigation salinity, bay design, for a given soil type. The model is a key component of a catchment salt export model under development by the CRC for Catchment Hydrology at Monash University.

Groundwater was observed to rise rapidly during an irrigation event. Within minutes the watertable was at the soil surface for the lighter soil type present at the field experiment. For the heavier clay soil in the bottom of the bay the watertable rise was much slower. For the event durations encountered, the watertable at the bottom of the bay usually failed to come to the soil surface. If however, the entire bay was composed of the lighter soil, the watertable would be at the surface throughout the bay, and exfiltration would occur.

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