SALT AND WATER MOVEMENT - MODELING

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ABSTRACT

Water and salt movement occurs under gradients from sources to sinks. This applies to the micro scale as well as the macro scale. Water and salt movement processes are responsible for much of the soil degradation that occurs.

Field assessment of water and salt movement includes the processes of infiltration, groundwater recharge, groundwater movement and groundwater discharge. After filling the available groundwater storage all recharge has to be balanced by groundwater discharge. The latter is in the form of seepage to creeks and rivers, capillary rise, artificial sub-surface drainage and uptake by roots of living plants.

Analytical solutions to groundwater and salt movement processes have been developed and are useful in many situations. Often, however, the complexity of the systems reduces the potential of such solutions, particularly where a larger scale is considered. Numerical groundwater flow models have overcome these problems to a degree, but there are still many uncertainties regarding the applicability of the results derived from these models.

The challenge for the future is to develop sound science-based models which link the flow processes of the unsaturated and the saturated zones and allow the assessment of solute transport, as well as the assessment of the proportion of the landscape affected by salt accumulation processes.

1. INTRODUCTION

The main processes of water and salt movement are infiltration, redistribution, percolation to the watertable, drainage, capillary rise, soil moisture extraction by roots and evaporation, (vertical) leakage between aquifers at different depths and groundwater flow. These processes are usually affected by changes in the hydrological cycle made by man, e.g. channel seepage, removal of trees and irrigation.

The salts in the landscape may originate from precipitation, weathering of deposited materials, transported in by river systems, or by groundwater flow. Quantification of these sources is necessary for a sound understanding of the salt dynamics in the landscape as a whole.

In general, groundwater flow occurs from sources to sinks under gradients. In some areas leaching predominates, whilst in other areas salt accumulation occurs.

2. THEORY OF WATER AND SALT MOVEMENT

2.1 Saturated Flow

The theory of saturated groundwater flow is based on the Darcy equation and the conservation of mass principle. For detailed discussion of theory reference is made to text books, such as Bear and Verruyt (1987) and Freeze and Cherry (1979).

Where the groundwater has varying density due to occurrence of brines, or where geothermal effects occur, the theory rapidly becomes more difficult.

Analytical solutions exist for many flow problems, but simplifications are usually necessary. This is often no problem where the scale considered is small, e.g. flow to tile drains, pumping from wells or seepage from channels. Where larger scales are involved however such simplifications tend to lead to generalisations, if not erroneous results. Computer technology has allowed the use of numerical solutions, which has proven useful in many situations.

2.2 Unsaturated Flow

Recent overviews of the theory of unsaturated flow are given by Nielsen *et al*. (1986) and Feddes *et al*. (1988). The soil water potential determines the unsaturated flow in soils. The soil moisture potential is the sum of individual potentials due to matric suction, gravity, gas pressure, osmotic effects and the overburden. In hydrological studies the potential is often expressed on a weight basis as a head h (m).

For field applications usually only the matric and gravity potentials are used, allowing the use of the Darcy equation with only one factor for the unsaturated hydraulic conductivity. The moisture flow is in the direction of lower potential.

Equations have been developed, mostly for isotropic, non-swelling soils, to describe changes in moisture level as a function of hydraulic conductivity, soil moisture potential gradients, time and elevation. A source/sink term may also be included. These equations and variants thereof are the basis for many analytical solutions of unsaturated flow, and are also the basis for numerical modelling.

The theory has been extended to the processes of infiltration, moisture redistribution, and capillary rise. Unfortunately field applicability of the theory is not simple. Hysteresis significantly affects the values of unsaturated hydraulic conductivity and diffusivity. In addition, there is usually a large spatial variation. This especially applies when more complex conditions are considered. Nielsen et al. (1986) and Feddes et al. (1988) give very good overviews of the effects of hysteresis, temperature, osmosis, and pH, electrolyte concentration, ionic composition, and swelling frozen soils.

2.3 Salt Movement Theory

The salt moving with the water is called the advective or convective component. The movement is subject to dispersion. Firstly there is the aspect of hydrodynamic dispersion. Secondly, along the path of flow there may be precipitation, due to chemical reactions with other salts, or due to pH, or redox potentials, or there may be interaction with the negative charge on the clay particles, causing the ions to move at a velocity different from the water. The effect may be different for each salt and soil condition. Bond and Smiles (1988) present equations by which the relative movement may be predicted.

Hydrodynamic dispersion is due to dispersion and molecular diffusion. Dispersion is the mechanical spreading in longitudinal and tranverse directions as the water flows through the pores. Molecular diffusion is significant only at very low flow rates, and would be considered only in cases of chemical contamination (Freeze and Cherry 1979). Models exist to calculate these features, however application in the field is tenuous as reliable information of dispersivities is hard to come by.

Chloride profile analysis may be used to assess the net downward movement. This has been successfully adopted for several situations, e.g. the Mallee in South Australia (Cook et al. 1988), and the dryland salting problems in Western Australia (Peck et al. 1981).

3. FIELD ASSESSMENT OF SALT AND WATER MOVEMENT

3.1 Infiltration and Redistribution

Infiltration into Riverina Plain soils is usually slow due to the presence of a (sodic) clay B-horizon or other less permeable zones. Cracking of clays has meant that often 30-100 mm infiltrates in a matter of minutes, and very little thereafter.

The tests for infiltration may involve the use of a ring inserted into the soil and ponding for a period, measuring the fall with time, or instruments capable of measuring sorptivity as well. Empirical or theoretical equations were used for analysis, e.g. the Philip equation or the Green and Ampt equation. The theory has been verified by field experiments and extended for two and three dimensional infiltration (e.g. Talsma 1970). The methodology is now developed to a stage where the infiltration process could be assessed at reasonable accuracy. However, an abundance of field data still does not exist, adding to the difficulties for modelers.

McIntyre et al. (1981), in a long term study of infiltration and redistribution in heavy clay soils at Benerembah, found that often the downward movement is throttled by a less permeable horizon. Below this layer the flow is unsaturated. After an initial decrease the rate stabilised to 0.9 mm/day for the Transitional Red Brown Earth for days 10-70, whilst for the swelling clay the rate stabilised at 3.5 mm/day.

Following infiltration to a deep watertable soil salt levels are usually low. During the initial stages of irrigation development leaching of salts is the dominating process. This situation is only reversed in parts of the landscape when a high watertable condition has been reached.

3.3 Irrigation Recharge

Before settlement in most of the Riverine Plain landscape recharge was very small to nil, except perhaps in flood years. The infiltrating water was used for plant transpiration, resulting in a salt bulge below the root zone. Clearing of hillsides and construction of irrigation schemes have resulted in more recharge, groundwater movement and solute transport.

Recharge under irrigation is from percolation from irrigated crops, rainfall, ponding of rice fields and channel seepage. Channel seepage may contribute up to 25% of all accessions (Gutteridge et al. 19856), although it needs to be noted that in many situations the figure is much less.

In Victoria perennial pastures may contribute 60-125 mm/yr to the ground water table on a sub-regional scale (Trewhella 1988), but much depends on the intensity of irrigation and the soil types. In wet winters rainfall tends to be the main contributor of accessions to the watertable, as shown by hydrographs. Nolan (1989) summarises various estimates of recharge in Victorian Districts.

Where rice is grown it is usually the main contributor for accessions. Up to 2000mm per season may percolate, with a regional average being about 400 mm in areas with deep watertables (e.g. van der Lelij 1987). Where the watertable is shallow the accessions are usually less than 100 mm, because the gradient for downward movement has become much less than unity.

In areas where rice growing is less intense, accessions from sources other than rice may dominate, as indicated by some preliminary model studies. For example, in the Berriquin Irrigation District the behaviour of hydrographs of bores shows distinct 'jumps' in wet years (Gutteridge _et al_.19856).

Recharge involves the displacement of occurring salts. For instance in the Coleambally Irrigation Area a downward shift in peak concentration of salts in aquifers of some 7 metres over 17 years may have occurred (van der Lelij, 1987). In the northern Victorian and Eastern NSW Irrigation Districts the charged shallow aquifers contain low salinity groundwater, which may be reused. To the western part of the plains, however, such downward leaching has never been of significance.

3.4 Groundwater Discharge

Groundwater discharge is in several forms. Some include the discharge of salts as well, others involve soil salt accumulation:

(i) Discharge to streams. Examples are the Barr Creek of Victoria, the Murray River downstream of Mildura and the Wakool River of NSW.

(ii) Discharge to constructed drains. This happens when the watertable level is above the bed of the drain, e.g. the Lalalty main drain near Finley.

(iii) Discharge through the soil, particularly in depressed areas. This process involves capillary rise and evaporation.

(iv) Discharge via roots of living plants.

(v) Discharge to artificial drains and vertical drainage bores.

Discharge processes need to balance the recharge once the storage factor (rising watertables) can no longer be used as a buffer of groundwater accessions.

Horizontal tile drainage has been installed in horticultural areas to control waterlogged conditions, both in the Shepparton area and in the M.I.A. Analytical solutions for drain spacing based on the Darcy equation and the Dupuit assumption have been used sucessfully.

Groundwater seepage into channels or seepage into a creek or river in many situations may be estimated from analytical solutions, provided permeability values and potentiometric information is available. These lower cost methods of analysis are quite satisfactory, especially where the results can be backed up by other calculations, e.g. the salt balance, or inflow/outflow measurements. The Barr Creek study by Gutteridge et al.19850) used a combination of water and salt balance and statistically based methods. In most more complex situations however it is probably preferable to use numerical techniques.

Drainage by pumping from aquifers can be achieved by design of appropriate tubewell or spearpoint systems. Pump testing of aquifers is often carried out for aquifer characterisation, but the resulting values for transmissivity and storativity usually are not sufficient as a guide for design purposes (van der Lelij 1978).

Discharge to or from deeper aquifers may be assessed from pumping tests using analysis for semi-confined conditions (Kruseman and de Ridder 1979). Unfortunately, results for the leakage factor are often erratic, because of aquifer and aquitard variation, and conditions during the test. Modeling, the other method of assessing deep leakage, also appears of doubtful reliability in this regard, but an order of magnitude improvement may nevertheless be achieved.

Trees as a dewatering technique may be quite effective. It is desirable that a small proportion (about 10%) of the groundwater that reaches the tree moves beyond the tree to avoid salinity buildup in the rootzone. This may be achieved by judicial choice of location. Unirrigated reserves are probably not the best choice.

Groundwater discharge to depressions is by means of capillary rise from a very shallow watertable. The upward movement may be calculated from the unsaturated hydraulic conductivity and soil moisture potential gradients. The unsaturated hydraulic conductivity as a function of soil moisture content decreases faster than the decrease in the value for the matric potential, having the effect that with infiltration into dry soil there is usually a distinct wetting front, and with capillary rise the upward movement virtually stops once the topsoil has been desiccated by evaporation processes in summer. When this happens the water movement is mainly in the vapour phase and affected by temperature gradients (Smiles 1977, van der Lelij 1983). Salt transport may be negligible.

With cracking clays being dessiccated, salt efflorescence has often been observed on the faces of the cracks. The inner part of the soil ped was less saline. Subsequent wetting by a rain shower would wash these salts to a deeper part of the soil crack.

The highest rates of upward movement (and salting) occur when the surface soil is moist and watertable conditions shallow. Watertables in rice growing areas are often raised over summer due to percolation, but discharge due to evaporation is small because of the presence of a very dry surface soil. In autumn the soil becomes moist due to rainfall and evaporation decreases, increasing capillary rise and upward salt transport. Discharge of groundwater and salinisation is season dependent.

Apart from seasonal factors the extraction of soil moisture by roots also effectively increases the potential gradients for capillary flow. It is known that irrigated crops effectively make use of a shallow non-saline watertable (Mason et al. 1983). A special study on soil cores (Percy et al. 1988) with a range of saline watertable conditions imposed and growing of wheat found a very significant increase in upward movement during the growing season, from 1 to 3 mm/day for the selfmulching clay soil, and from 0.3 to 1.5 mm/day for the Transitional Red Brown Earth (watertable at 0.85 metres).

Within a soil association there is spatial variation in infiltration, but also in capillary rise. The patterns of up and downward movement are unlikely to coincide, and in a high watertable area there are likely to be salt accumulation processes taking place only metres away from where leaching occurs.

3.5 Groundwater Movement

Groundwater movement is from a recharge area (source) to a discharge area (sink). The times of travel may be short, as with seepage from channels to adjacent land, to thousands of years through deeper aquifers.

Studies of groundwater movement are usually based on observations on networks of piezometers installed in aquifer systems at different depths, groundwater contour maps, rate of rise maps, and examination of bore and salinity profiles and land use data. By close scrutiny of all the data at hand the processes at work can be evaluated. Quantification is only possible with additional information, such as pump test data for transmissivity and leakage factors, groundwater pump rates and recharge assessments for different cropping systems.

Analytical models which exist to describe flow in Riverine Plain conditions include sub-surface drainage applications, aquifer evaluations, seepage from and to channels and flow from rice fields to adjacent land (van der Lelij 1980). For quantification of complex systems the use of numerical models is usually necessary.

4. GROUNDWATER MODELING

Modeling is an attempt to duplicate the behaviour of natural systems by using the laws of science and mathematics. A natural system can be represented by a set of equations which may be solved by a computer. Some of the equations used may only be based on assumptions about the factors involved and therefore represent an approximation of the actual situation. There will always be gaps in the knowledge of data which need to be bridged. The goodness of a model therefore needs to be judged on the assumptions and the reliability of the critical data.

Soil water and salt movement models may be separated into unsaturated and saturated flow problems.

4.1 Saturated Groundwater Flow

Numerical modeling helps in understanding the groundwater system, it is a means of synthesising field data into a consistent framework, aquifer behaviour may be evaluated and if a simulation can be achieved, predictions may be made about behaviour for different inputs.

With saturated flow of aquifers a great many models exist, with many solution techniques. For discretisation most models use either finite element or finite difference techniques, but the boundary integral equation method is also used (Bear and Verruyt 1987).

Finite differences models involve the superposition of a grid of rectangular cells over the area of interest. Data are compiled for each cell. Finite element involves the setting of nodes for which data are available and the creation of a set of polygons (usually triangles) to connect the nodes.

In finite differences models each cell has a unique equation solving the hydraulic head from the data, which consist of aquifer properties, recharge and discharge. Simultaneous equations are solved by iteration for which there are several methods. For further details see (Bear and Verruyt 1987).

An overview of groundwater models is given by (Konikow and Mercer 1988) who state that the mathematics of modeling are no longer the constraint on further progress. The paucity of data for the area in which the model is used, or poorly defined boundary conditions are usually the main limitation. Because of the varying velocity between water and solute molecules in the system (section above) to date only two solute transport models have been developed.

4.1 Unsaturated Flow

Analytical models exist to describe the water and salt transport in the unsaturated zone. For instance Rose et al. (1980) distinguish spatially-lumped steady state models to estimate the leaching requirement in a well drained soil to maintain a certain water salinity at the bottom of the rootzone, and a lumped parameter non-steady state model in slowly draining soils to predict the soil salinity at a depth after a long period of near constant percolation. Peck et al. (1981) describe the use of models in W.A. regions to analyse chloride profiles for leaching effects of changed hydrology of catchments.

Numerical unsaturated flow models are usually based on the Richards equation (Bear and Verruyt, 1987). The soil is divided in sufficiently thin layers, or grid cells, to allow for assessment of factors such as unsaturated hydraulic conductivity and diffusivity from soil moisture values in each.

4.2 Composite Groundwater Models

Linkage of groundwater numerical models with recharge and discharge features allows the analysis of complex systems. The conceptualisation of the processes is crucial. Often the complex theory of infiltration, recharge, capillary rise and evaporation are estimated by empirical process functions, put together in a so-called process model. This process model may be linked with a two-dimensional groundwater flow model, giving a three-dimensional simulation.

Data for some of the factors may not be available and has to be estimated, or "modeled". This may or may not be a problem, dependent on the overall impact of the particular process on the rest of the model. Sensitivity analysis therefore is an important part of the process of developing these models. The modeling process often is time consuming and tedious.

Lack of availability of input data are particularly relevant to the Riverine Plain. For instance with a model developed for the Berriquin Irrigation District the aim was to combine surface hydrology features such as irrigation, rainfall and runoff with recharge and groundwater flow, to calculate the effect of watertable behaviour on seepage into surface drains. A cell size of 2.5m x 2.5 km and time steps of a month or more were used. The data averaging for each cell for soil characteristics, infiltration, surface level, irrigation recharge rates etc. made it very difficult to derive meaningful conclusions. The model could be assessed as being too ambitious.

The interaction between factors may cause special difficulties for the calibration process, e.g. to estimate recharge from water level changes the storativity factor needs to be known fairly accurately, and this often is not the case. There are other such cross linkages.

Numerical models greatly help in the analysis of complex flow problems, particularly on a larger scale. The challenge for the future is to develop better science-based models to determine land and water management strategies for the irrigated plains environment. These models would implicitly link the flow processes of the unsaturated and the saturated zones, allow the assessment of solute transport, and the assessment of the proportion of the landscape affected by salt accumulation processes.

5. REFERENCES

Bear, J. and Verruyt (1987) The Hydraulics of Groundwater. McGraw-Hill, New York. 567pp.

Bond, W.J. and Smiles, D.E. (1988) Predicting the average movement of reactive solutes in soils. Soil Use Water Manage. 4:(4) 115-20.

Cook, P.G. Jolly, I.D. Walker, G.R. and Walker, C.D. (1988) Localised recharge in the Western Murray Basin. Murray Basin 88: Abstracts of Geol. Grw. and Salinity Conf. Canberra. May 1988.pp.43-45.

Feddes, R.A. Kabat, P. van Bakel, P.J.T Bronswijk, J.J.B and Halbertsma, J. (1988) Modeling soil water dynamics in the unsaturated zone - state of the art. J. Hydrol. 100: 69-111.

Freeze, R.A. and Cherry, J.A. (1979) Groundwater. Prentice Hall Inc., Englewood, N.J. 07632.

Gutteridge, Haskins and Davey, ACIL Aust and Naunton (1985a). Study of Barr Creek and Catchment. Report RWC of Victoria, July 1985.

Gutteridge, Haskins and Davey (1985b) Study of salinity and waterlogging in irrigated areas of N.S.W. N.S.W. Dept. of Water Resources Report.

Konikow, L.F. and Mercer, J.W. (1988) Groundwater flow and transport modelling. J. Hydrol, 100, 379-409.

Kruseman, G.P. and de Ridder, N.A. (1979) Analysis and Evaluation of pump test data. Veenman, Wageningen, The Netherlands.

Mason, W.K. Meyer, W.S. Smith, R.C.G. and Barrs, H.D.(1983) Water balance of three irrigated crops on fine textured soils of the Riverine Plain. Aust J. Agric. Res. 34: 183-91.

McIntyre, D.S. Loveday, J. and Watson, C.L. (1981) Field studies of water and salt movement in an irrigated swelling clay soil. Infiltration during ponding. Aust J.Soil Res. 20: 81 - 90.

Nielsen, D.R. van Genuchten, M.Th. and Biggar, J.W.(1986) Water and solute transport processes in the unsaturated zone. Water Resources Res. 22:(9) 89S-108S.

Peck, A.J. Johnston, C.D. and Williamson, D.R. (1981) Analysis of solute distributions in deeply weathered soils. Proc. Dry Land Salinity Conf., Perth, WA. 6.1-6.9

Percy, H.P., van der Lelij, A., and Meyer, W.S. (1988) Salt Accumulation in Soils Subjected to High Watertables. Dept. Water Resources. Griffith, NSW.

Rose, C.W. Parlange, J.Y. and Saffigna, P.G. (1980) Transport Processes in Salinisation. School Aust Env. Studies, Griffith Uni.Qld.

Smiles, D.E. (1977) Air water heat relationships. In: Soil Factors, Eds. Russell, J.S. and Graecen, E.L.

Talsma, T,(1970). Some aspects of three dimensional infiltration. Aust. J. Soil Res. 8: 179-84.

Talsma, T. (1985) Prediction of hydraulic conductivity data from soil water retention data. Soil Sci. 140 184-88.

Trewhella, W. (1988) Irrigation recharge. In: Murray Basin 88: Abstracts of Geol. Grw. and Salinity Conf. Canberra, May 1988, 169-71.

van der Lelij, A. (1978) Artificial sub-surface drainage in N.S.W. In: The Hydrologeology of the Riverine Plain of South East - Australia. Aust. Soc. Soil Sci.Inc.,Riverina Branch, Griffith, NSW, 209-28.

van der Lelij, A. (1980) Groundwater flow from rice field to adjacent land. Proc. Salinity and Water Quality Conf., Toowoomba, Qld, 211-18.

van der Lelij, A. (1983) Moisture loss from clay soils subjected to high watertable conditions - A discussion based on a preliminary drum experiment.Soils Conference Proc. Aust. Soc. Soil Sci. Inc., Riverina Branch, Yanco, May 83, 79-86.

van der Lelij, A. (1987) Groundwater accessions in the Coleambally Irrigation Area. In: Groundwater Recharge, Ed. M.L. Sharma. Proc. Grw. Rech. Conf, Mandurah, W.A. 281-90.