1 Water Balance Modelling: Concepts and Applications

Lu Zhang,* Glen R. Walker† and Warrick R. Dawes*

Abstract

Many environmental problems are caused by changes in aspects of the hydrological cycle. Water balance modelling combined with field experiments can give us a better understanding of the components of the hydrological cycle from which to develop appropriate management options. Water balance models can be constructed at any level of complexity. In simple 'bucket' models only the most important processes are represented. When appropriately used, bucket models can provide useful insights into the functional behaviour of a system. Complex models are needed to understand complex feedbacks and interactions among different processes of the system. However, increasing the complexity of a model does not necessarily lead to a more accurate model and it is essential that model complexity matches the availability of data. The key to successful water balance modelling is to have a clearly defined objective and to select an appropriate model. This chapter outlines the principles of water balance modelling and explains how models can be used in crop management.

水文循环要素的变化引起了许多环境问题。水量平衡模型结合野外试验，可以更好地了解水文循环各组成成分、选择合适的管理方式。水量平衡模型可以建造在任意复杂的程度上。简单的“桶式”模型只考虑了最重要的水文循环过程，有助于了解系统内部的功能。如果要理解系统内部过程间的复杂反馈以及相互作用，则需要复杂的模型。不过，增加一个模型的复杂性并不一定就增加了它的精确性，模型的复杂程度必须与能得到的数据相匹配。有一个明确的目标，选择一个合适的模型是建模成功的关键。本文概述了水量平衡模型化的原则，并说明了如何将模型应用在农作物管理方面。

* CSIRO Land and Water, PO Box 1666, Canberra, ACT 2601, Australia. Email: lu.zhang@csiro.au

† CSIRO Land and Water, Adelaide, SA 5064, Australia.

Lu Zhang, Walker, G.R. and Dawes, W.R. 2002. Water balance modelling: concepts and applications. In: McVicar, T.R., Li Rui, Walker, J., Fitzpatrick, R.W. and Liu Changming (eds), *Regional Water and Soil Assessment for Managing Sustainable Agriculture in China and Australia*, ACIAR Monograph No. 84, 31–47.

THE MOVEMENT of water through the continuum of the soil, vegetation and atmosphere is an important process central to energy, carbon and solute balances. The system is integrated, so changes in one part of the system will affect the others and we need to consider the dynamic interactions and feedback between the processes.

Most current environmental problems arise from tampering with one, or a few, aspects of the system with no understanding of the function of the system as a whole. In Australia, much environmental degradation, including salinisation, is associated with changes in the near-surface water balance induced by massive clearing of native vegetation. These changes have led to significant increases in groundwater recharge, which in turn have led to rising watertables and salinisation. In contrast, the groundwater tables on the North China Plain (NCP) have declined significantly due to overuse of groundwater for irrigation. This presents a serious problem for sustainable agricultural development in the region because there is no reliable surface water for irrigation.

Field techniques alone cannot identify optimal or appropriate land use, because changes occur over a large area and a long time. Also, field experiments are expensive and only a limited number of land-use options can be trialled. For similar reasons, there is a limit to the range of soils and catchments that can be used for field trials. One way to overcome the problem is to combine field experiments with water balance modelling. Modelling takes account of climate variability because it can be used to objectively analyse climatic data and to extrapolate results from short-term field trials over periods of many years. It can also provide an insight into feedbacks and linkages, and help set priorities for field experimentation and data collection.

This chapter deals with the concepts of surface water balance modelling and the processes involved. It briefly describes the WAVES (water, atmosphere, vegetation, energy, soil) model and its application to the NCP and the Australian Mallee region.

Understanding the Hydrological Cycle

The natural occurrence of water circulation near the surface of the Earth, the hydrological cycle, is illustrated in Figure 1. Radiant energy from the sun is the driving force of the cycle, which is generally considered to begin with the evaporation of water from oceans. The resulting water vapour is transported to the atmosphere, where, with favourable conditions, it may produce precipitation. Precipitation may:

- be intercepted by vegetation and evaporated directly back to the atmosphere;
- infiltrate the soil and evaporate from the soil surface or be transpired by vegetation;
- become surface runoff; or
- drain through the soil to form groundwater recharge.

Vegetation influences the hydrological cycle through the exchange of energy, water, carbon and other substances and is therefore critical for many hydrological processes, in particular transpiration, infiltration and runoff. The land phase of the hydrological cycle is of particular interest to those studying the problem of deep drainage.

The movement of water through the hydrological cycle varies significantly in both time and space. Australia, the driest continent, has the highest variability in rainfall and runoff and is therefore a difficult system to model. Torrential rains can flood even the driest part of the continent, while in adjacent areas the story may be very different. The hydrological cycle emphasises the four phases of interest to hydrologists: precipitation, evapotranspiration, surface runoff and groundwater. However, in the context of dryland salinity and recharge control, it is also important to

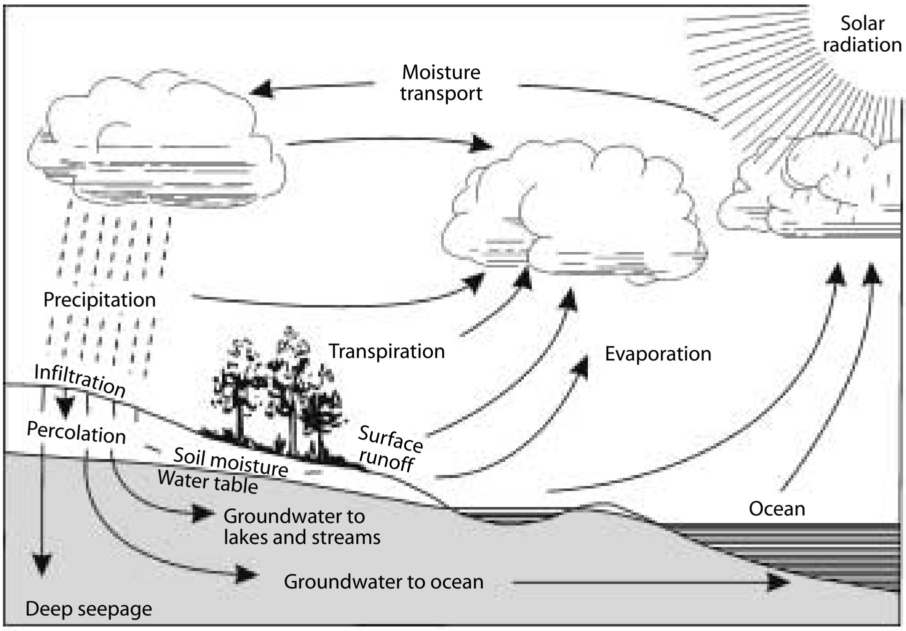

Figure 1. The hydrological cycle.

determine deep drainage and changes in soil water storage. In addition, there can be different relationships between surface water and groundwater. For example, a stream or stream reach can receive groundwater as baseflow or discharge to groundwater. Understanding these relationships can help us to estimate catchment-scale water balance.

Figure 2 shows the three components of the runoff from a catchment:

- surface runoff or overland flow (Q_1)
- subsurface runoff or interflow (Q_2)
- groundwater runoff or baseflow (Q_3).

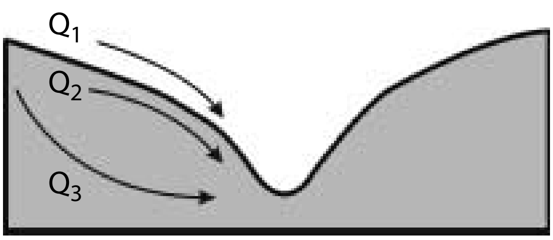

Figure 2. Runoff components.

Surface runoff or overland flow is the water that travels over the surface of the ground towards the stream channel. It can be generated by two mechanisms: infiltration excess runoff (Hortonian overland flow) and saturation excess runoff (Dune and Black overland flow) (Freeze 1974). Infiltration excess runoff often occurs from patches where soils become saturated at the surface. This happens when rainfall intensity exceeds the infiltration capacity of the soils. Hortonian overland flow is an important runoff mechanism in arid and semiarid regions, where rainfalls tend to be intensive and surface infiltration rates low. Saturation excess runoff is generated by rainfall on areas where the soil is already saturated from below. This mechanism of surface runoff generation occurs primarily on the lower slopes of the catchment and along valley bottoms adjacent to stream channels.

Subsurface runoff or interflow represents that portion of infiltrated rainfall that moves laterally through the upper soil layers until it reaches the stream channel. Subsurface runoff moves more slowly than surface runoff. The proportion of total runoff that occurs as subsurface runoff or interflow depends on space–time properties of rainfall and physical characteristics of the catchment. A thin soil layer overlying more impermeable soil layers tends to promote subsurface runoff or interflow, whereas uniformly permeable soil encourages downward movement of infiltrated rainfall.

Subsurface runoff has always been a popular concept amongst forest hydrologists — indeed amongst humid temperate hydrologists generally. Dunne (1978) termed it subsurface stormflow and used physical models and hydrograph data to show that, even on small catchments, this flow peaked some hours, or even days, after the cessation of rainfall. The real problem is how to model water movement laterally through the soil on hillslopes down to the streamline in real soils. Because this principally occurs in forest with deep litter layers, it has been suggested that most water moves through this litter (see Dunne 1983). However, Beven (1981) has suggested that in fact extensive exfiltration at low rates can create a very shallow flow moving at reasonable speed that meets the time requirements. This is particularly the case where topographic convergence concentrates both the subsurface and exfiltrated water.

Thus, in sloping situations, particularly if there is a reduction in permeability with depth, lateral flow can develop in both the litter layer and the upper soil zone. Water then flows laterally downslope, mostly in a temporarily saturated zone, giving rise to the streamflow element termed interflow or intermediate response flow. The flow emerges in lower parts of the landscape or in the banks of channels and streams to contribute to the streamflow as interflow.

Groundwater runoff, or baseflow, is that portion of infiltrated rainfall that reaches watertables and then discharges into streams. It responds much more slowly to rainfall and does not fluctuate rapidly. It represents the drainage of water to the streamline from the regional or deep groundwater, or both. A minimum catchment area is often needed for such a response to be evident or for the regional groundwater surface to reach the stream incision level. Where we have seasonal rainfall we find that there is virtually an annual hydrograph of baseflow, with levels building up and peaking towards the end of the wet season. If the total outflow is low, then the baseflow may be intermittent or seasonal.

It should be clear that the distinctions between the three components of total runoff are arbitrary. These components can occur separately or simultaneously with varying magnitudes, depending on a combination of climatic and physical conditions of the catchment.

Water Balance

Water balance is based on the law of conservation of mass: any change in the water content of a given soil volume during a specified period must equal the difference between the amount of water added to the soil volume and the amount of water withdrawn from it. In other words, the water content of the soil volume will increase when additional water from outside is added by infiltration or capillary rise, and decrease when water is withdrawn by evapotranspiration or deep drainage. The control soil volume for which the water balance is computed is often determined arbitrarily. In principle, a water balance can be computed for any soil volume, ranging from a small sample of soil to an entire catchment. For the purpose of recharge estimation, it is generally appropriate to consider the root zone as the control volume and express the water balance per unit area.

Figure 3 depicts the water balance for a root zone. The processes represented form a part of the overall hydrological cycle (Fig. 1). Much of the rain that falls during the first part of a storm is intercepted by the canopy and evaporates directly back to the atmosphere. Rainfall that reaches the ground may either infiltrate the soil or run off. Some of the water that infiltrates the soil evaporates directly from the soil surface or is transpired by plants; some may be stored in the soil profile. Water that moves laterally across the B horizon of the soil is known as subsurface flow; water that moves vertically is known as deep drainage. Deep drainage should not be confused with infiltration. Infiltration is the amount of water that enters the soil; deep drainage is the flux of water below the rooting depth.

The root zone water balance is usually expressed as:

$$\Delta S = P - I - E - T - RO - DD \qquad (1)$$

where ΔS is the change in root zone soil water storage over the time period of interest, P is precipitation, I is interception loss, E is direct evaporation from the soil surface, T is transpiration by plants, RO is surface runoff or overland flow, and DD is deep drainage out of the root zone. All quantities are expressed in terms of volume of water per unit land area or equivalent depth of water over the period considered.

The recharge to the groundwater system can be calculated as:

$$R = DD - SSF \qquad (2)$$

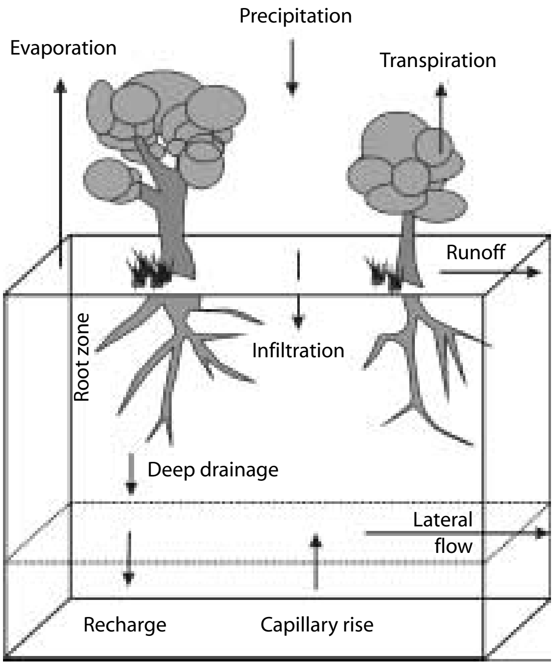

Figure 3. Schematic diagram of the water balance for a root zone.

where SSF is the lateral subsurface flow calculated according to Darcy's law (see 'WAVES' section below).

When the control volume is the entire catchment (Fig. 4), the surface water balance equation can be expressed as:

$$\Delta\langle S\rangle = \langle P\rangle - \langle ET\rangle - \langle Q\rangle - \langle R\rangle \qquad (3)$$

where $\Delta\langle S\rangle$ is the change in spatially averaged catchment water storage, $\langle P\rangle$ is the spatially averaged precipitation, $\langle ET\rangle$ is the spatially averaged evapotranspiration, $\langle Q\rangle$ is the spatially averaged catchment surface runoff, and $\langle R\rangle$ is the spatially averaged catchment recharge.

The root zone water balance shown in Figure 3 can be considered as a plot-sized profile in a catchment, as shown in Figure 4. It is tempting to think of the catchment as a collection of such plots, with recharge estimates from each plot being simply added to yield the total recharge for the catchment. A few catchments do operate this way, but most exhibit complex lateral redistribution of water, so it is difficult to estimate catchment scale recharge. One should not assume that recharge estimates from a plot-scale water balance equal the catchment-scale estimates unless thorough hydrogeological investigations are undertaken: plots and catchments differ in terms of hydrological processes, recharge pathways and spatial heterogeneity of soil properties.

Evaluation of Water Balance

The root zone water balance presented by equations (1) and (2) is the basis of water balance modelling. The main advantages of the method are that it uses available data (rainfall, runoff) and has a clear conceptual basis. It seems simple in principle, but in practice it is difficult to measure or estimate each of the components. The evaluation of the water balance equation requires information about the system considered and adequate data.

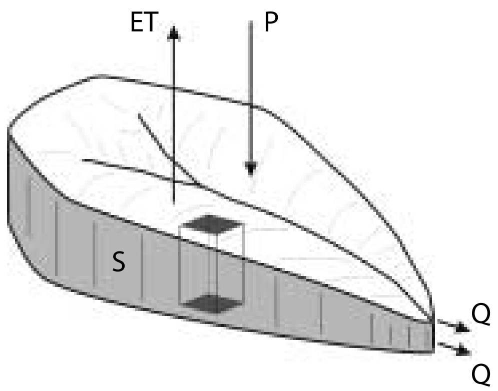

Figure 4. Schematic diagram of a catchment. The box indicates the control volume shown in Figure 3. S = water balance; ET = evapotranspiration; P = precipitation; Q = water runoff or water flow.

Precipitation is often the largest term in the water balance equation and can be measured directly using rain gauges. Interception loss is a complex process affected by factors such as rainfall regime and canopy characteristics. Interception loss can be measured directly in the field or estimated using the method of Horton (1919). Soil evaporation is often lumped together with plant transpiration as total evapotranspiration, which forms the second or third largest term in the water balance equation. Evapotranspiration can be estimated from meteorological and soil moisture data or measured directly. In agricultural fields, the amount of surface runoff generally is considered negligible. However, at catchment scale, runoff may be significant compared to the major components of the water balance; evaluation of this term can be difficult due to different runoff components. The evaluation of the storage term depends on the time period over which the water balance is computed. On an annual basis, the change in water content of the root zone is likely to be small in relation to the total water balance and can be neglected. Over a shorter period, the change in the soil water storage can be significant and must be considered. Different techniques such as the neutron meter and time domain reflectometry (TDR) can be used to measure soil water content. Deep drainage is often only a small fraction of the precipitation — 5% is a typical figure.

Water Balance Modelling

A water balance model can be considered as a system of equations designed to represent some aspects of the hydrological cycle. Depending on the objectives of the study and data availability, modelling can have different levels of complexity, although the model is a simplification of the real world, no matter how complex it may be. A simple bucket model may be suitable for some purposes; in other cases more complex models may be required. It is important to recognise that increasing model complexity does not necessarily improve accuracy (Walker and Zhang 2001).

Simple bucket models

Conceptualisation of the system is based on our understanding of the hydrological cycle and the different pathways that water can take within a catchment. In the simplest case, the control volume (e.g. paddock) is considered as a bucket that is filled up by rainfall and emptied by evapotranspiration. When the bucket is full, extra water is assumed to become deep drainage. The only input data required by this model are rainfall, actual evapotranspiration estimated from potential evapotranspiration and soil water content, and the available water storage capacity. An example of such a model is shown in Walker and Zhang (2001). There are a number of variations to the simple bucket model depending on conceptualisation of the system and methods for estimating evapotranspiration (Walker and Zhang 2001; White et al. 2000; Sophocleous 1991; Rowell 1994; Lerner et al. 1990; Scotter et al. 1979).

Complex models

More complex models are available that deal with soil moisture dynamics — feedback between plant growth and soil moisture (Walker and Zhang 2001). These models are designed to simulate complex interactions within the system, to explore sensitivities to different assumptions and to provide more rigorous analysis of experimental results.

Most have been reasonably well documented and people can be trained to use the models (Walker and Zhang 2001).

However, there are tradeoffs in choosing between complex models and the simple bucket models. One of the drawbacks in using more complex models is that they require more data and more time on the part of the user to understand them. Using a complex model without understanding its general structure, parameter space and input variables can cause problems, and interpreting results from such a model can be difficult because of feedback between processes. There is little point in selecting a complex model if sufficient data are not available. The key to successful modelling is to have very clear objectives, a good understanding of the system and a clear identification of appropriate representations (i.e. a clear conceptualisation of the system and matching of model complexity with data availability). Practical issues associated with complex models are discussed in detail by Hook et al. (1998) and Walker and Zhang (2001).

Matching model complexity with data availability

The present array of different water balance models exists because model users have different applications and purposes, and because of the variety in landscapes and climatic conditions. For some purposes, very simple single bucket water balance models are suitable, but other uses require greater functionality—for example, the ability to predict plant growth and grain yield. With greater functionality comes greater complexity. More complex models often require greater effort in parameterisation, more computing power, and additional work in interpreting results. Perhaps more significant is that error propagation can be more difficult to understand and detect. When selecting a model for a particular application a user needs to balance the desired functionality against complexity and data requirements.

The purpose of water balance modelling is generally to improve our understanding of the critical processes that influence the hydrological cycle and to extend knowledge from field or laboratory experiments to quantitative predictions for other sites and climates. The models are always a simplification of the field processes, but they attempt to account for the most important factors that influence the water balance. Adding predictions of less important processes might give diminishing returns if the costs in greater model complexity start to outweigh the extra value from additional functionality. Different models are just different balances between functionality and complexity; users need to chose a tool with an appropriate balance for their particular purpose. In general, a good rule of thumb is to avoid unnecessary complexity and to achieve a level of process detail consistent with the importance of the process for the application in question. There is not much point in having relatively unimportant processes represented in great detail.

In achieving a balance between simplicity and complexity, users should be aware of two types of modelling errors. The first is ‘systematic error’ resulting from simplifying assumptions (for example, not considering runoff or macropore flow). As we add more processes to the model and increase its complexity (see Fig. 5), we generally decrease the systematic error. The second type of error is ‘calibration error’ and results from our lack of knowledge of the parameters that are needed for the model. Generally, as complexity increases, there is greater risk of parameterisation error increasing. There is a need to reach some type of balance. However, it is not easy to quantify the systematic error, so it is not always easy to define exactly the balance between simplicity and complexity. As a guiding principle, a relatively simple model is likely to be required if there are limited data. The simplest model that we can usefully apply is one that captures only those factors that are critical to the processes that are being investigated.

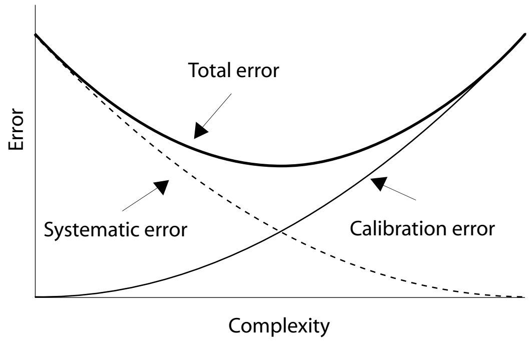

Figure 5. Schematic diagram showing systematic and calibration error.

Model parameterisation

With a significant increase in model complexity there is generally an increase in the number of parameters that are needed. Rainfall and evapotranspiration data are often available. The use of measured data to parameterise models (*direct estimation)* is ideal, but is not always possible. For example, detailed measurements of saturated or unsaturated hydraulic conductivity are usually not available away from research sites. Moreover, we often want to infer measurements over large areas. Where do we obtain these parameters?

One approach is to use more readily available information to estimate the attributes of interest. This is known as *knowledge-based estimation.* For example, there are predictive relationships between soil texture and a range of soil hydraulic properties. These can be combined with information on the distribution of different soil types to make spatial predictions of soil properties. So there is a range of methods for predicting well-defined and measurable *physical parameters*. However, some models contain *process parameters* that are artefacts of the model and not quantities that can be directly measured. Examples of *physical parameters* include water content and saturated hydraulic conductivity; *process parameters* include effective roughness, leaf mortality coefficients, or coefficients that control the rate of water movement between layers in a tipping bucket model. Such *process parameters* can be determined only through *model calibration*.

Model calibration is the process of inferring input parameter values by fitting model outputs to a set of detailed measurements. This constitutes the careful tuning of a model so that its output matches measurements. This should be avoided where possible as it is less desirable than either direct measurement or knowledge-based estimation. If calibration is necessary, the number of parameters that are calibrated should be kept to a minimum and researchers should ensure that the exercise does not result in the model getting the right answers (i.e. matching the measured data) for the wrong reasons. An incorrect mix of input parameter values might enable adequate matching of one set of data but the process description might still be inadequate and fail completely when applied in another set of conditions. Error due to poorly fitted parameters is called *calibration error.* It is desirable in any calibration to consider the function of different submodels separately, rather than just calibrating against integral model outputs such as profile water content.

Generally, models are parameterised using a combination of direct measurement, knowledge-based estimation and a minimum of calibration. Available data for parameterisation are often at the wrong spatial or time scale and sometimes inputs are highly correlated. It is important to have a good knowledge of the sensitivity of the model outputs to the parameterisation data. Then the user will know where to put the most effort into parameterisation through direct measurement, and will have a better understanding of the potential errors in model predictions.

Model testing and sensitivity analysis

Model testing is an essential step in any model development. In a strict sense, no model can be validated. Models are a simplification of reality, so it is necessary to build assumptions into the model. However, the more consistent a tested model is with measured information, the greater the confidence we may have in the tested model. If one is using the model as an educational or explorative tool, it may not matter that the parameter values do not match reality exactly, as long as they are approximately right. However, if we are using the model for predictions, we must have confidence in the key parameters (Walker and Zhang 2001).

For this reason, we must understand how the conclusions relate to the assumptions of the model and the data used in the model parameterisation. A tested model that is useful for a given objective is one in which the conclusions are robust to both the assumptions and the data. Testing of a model involves a number of steps:

- testing whether assumptions are reasonable;
- testing whether code matches conceptualisation;
- testing the sensitivity of the model to input data and model parameters; and
- testing model outputs against observed data that were not used in model parameterisation.

In going through these steps, it is clear that model testing will always be partly subjective. The simplest method of sensitivity analysis is to vary input parameter values by a set amount or percentage, and evaluate the resulting changes in model output. This provides information about the propagation of error from input data to conclusions. If parameters are changed independently, some important interactions between parameters may be missed. These interactions occur when the response of the model to a change in a particular parameter depends on the values chosen for other parameters. The choice of output variable to use as a measure of model sensitivity is not trivial. Conclusions about the sensitivity or uncertainty analysis will usually change depending on which output parameter was chosen, whether the average, maximum or minimum was chosen, and which spatial and temporal condition was chosen.

WAVES—An Integrated One-Dimensional Energy and Water Balance Model

The WAVES model is designed to simulate energy, water, carbon and solute balances of a one-dimensional soil–canopy–atmosphere system (Dawes and Short 1993; Zhang et al. 1996). It is a process-based model that integrates soil and canopy–atmosphere with a consistent level of process detail. WAVES predicts the dynamic interactions and feedbacks between the processes. Thus, the model is well suited to investigations of hydrological and ecological responses to changes in land management and weather, such as those discussed above.

WAVES models the following processes on a daily time step:

- interception of rainfall and light by the canopy;
- surface energy balance;
- carbon balance and plant growth;
- soil evaporation and canopy evapotranspiration;
- surface runoff and infiltration;
- saturated/unsaturated soil moisture dynamics (soil water content with depth);
- drainage (recharge);
- solute transport of salt (NaCl); and
- watertable interactions.

The WAVES model is based on five balances:

- ***energy balance**:* partitions available energy into canopy and soil for plant growth and evapotranspiration (Beer’s law);
- ***water balance**:* handles infiltration, runoff, evapotranspiration (Penman–Monteith equation), soil moisture redistribution

(Richards equation), drainage, and watertable interactions;

- ***carbon balance****:* calculates carbon assimilation using integrated rate methodology (IRM), dynamically allocates carbon to leaves, stems and roots, and estimates canopy resistance for plant transpiration;
- ***solute balance***: estimates conservative solute transport within the soil column and the impact of salinity on plants (osmotic effect only); and
- ***balance*** of complexity, usefulness, and accuracy.

Energy balance

The energy balance module calculates net radiation from incoming solar radiation, air temperature and humidity, then partitions it into canopy and soil available energy using Beer's law. Evapotranspiration is calculated using the Penman–Monteith equation (Monteith 1981) with available energy, vapour pressure deficit, and air temperature as inputs. The Penman–Monteith equation is a 'big leaf' model based on the combination of energy balance and aerodynamic principles. It requires estimation of aerodynamic and canopy resistances. The aerodynamic resistance is estimated from wind speed and surface roughness, while canopy resistance is calculated as a function of net assimilation rate, vapour pressure deficit, and carbon dioxide concentration. WAVES couples canopy and atmospheric data using the approach proposed by Jarvis and McNaughton (1986); it handles multilayer canopies explicitly.

WAVES assumes that the canopy and ground surface temperatures are equal to the average daily air temperature. This assumption does not introduce much error into the energy balance for relatively dense plant stands with nonlimiting water supply (Zhang et al. 1996). The ground heat flux is neglected in the energy balance equation because over land surfaces the daily mean value of the ground heat flux is one or more orders of magnitude smaller than the net radiation.

Water balance

The soil water balance module handles rainfall infiltration, overland flow, soil and plant water extraction, moisture redistribution, drainage (recharge), and watertable interactions. Soil water movement in both the unsaturated and saturated zones is simulated using a fully implicit finite-difference numerical solution of a mixed form of Richards' equation (Richards 1931; Dawes and Short 1993; Short et al. 1995). A full description of the solution to Richards' equation can be found in Dawes and Short (1993). Overland flow can be generated when the rainfall rate exceeds the infiltration rate of the soil, and when rain falls on a saturated surface. Both of the mechanisms are considered explicitly in WAVES. A watertable may develop anywhere within the soil profile. If nonzero slope is specified as input, then lateral subsurface flow occurs via any saturated watertable at a soil layer boundary, and is described by Darcy's law. Researchers can specify a regional groundwater depth, which may be changed daily according to changes in weather, and which can be used to interact with the WAVES soil column. Evaporation and transpiration draw water out of the soil; when the internal saturated water level is below the regional watertable, leakage into the column occurs and may bring salt with it. Conversely, when the internal water level is above the regional watertable, due to plant inactivity or large amounts of infiltration, water may leak out of the column and leach salts.

To solve Richards' equation, the analytical soil model of Broadbridge and White (1988) is used to describe the relationships among water potential, volumetric water content and hydraulic conductivity. This soil model has five parameters: saturated hydraulic conductivity, volumetric soil moisture content at saturation, air-dry volumetric water content, soil capillary length scale, and a soil structure parameter. The Broadbridge and White (1988) soil model can realistically represent a comprehensive range of soil moisture characteristics, from highly nonlinear associated with a well-developed capillary fringe, to weakly nonlinear associated with highly structured soil and macropores.

The assumptions of Richards' equation are that the soil is incompressible, non-hysteretic and isothermal, and that moisture moves in a single phase only. The equation also assumes that flow is via the soil matrix only, and not via macropores and larger preferred pathways. The soil is assumed to be isotropic for the formulation of Darcy's equation for lateral movement. Any water ponded on the surface can either be left to pond, or appear as runoff within the time step. Soil air flow is ignored.

Carbon balance

The carbon balance and plant growth module calculates daily carbon assimilation from a maximum value and the relative availability of light, water and nutrients. The limiting effects of temperature and salt in the soil water on assimilation are modelled explicitly. It is assumed that the actual growth rate depends on the potential growth rate and the level of the available resources. To combine the three limiting factors on plant growth into a single scalar, we use the IRM of Wu et al. (1994), which allows other limiting factors, such as atmospheric carbon dioxide concentration, to be easily included. Once carbon assimilation is calculated, it is used as input to the dynamic allocation of carbon to leaves, stems and roots, and into the calculation of canopy resistance to transpiration.

Solute balance

Solute transport within the soil column is solved with a convection–dispersion equation, in the same way as soil moisture dynamics (Dawes and Short 1993). It is assumed that the solute concentration does not interact with soil hydraulic properties, so water fluxes and contents are constants with respect to the solutes, and that salt never crystallises out of solution. This makes the solution of solute dynamics explicit. The feedback of salinity to plants is through the reduction in apparent available water due to the osmotic potential induced by dissolved salt (NaCl) alone.

Overall model balance

WAVES emphasises the physical aspects of soil water fluxes and the physiological control of water loss through transpiration. It can be used to simulate the hydrological and ecological effects of scenario management options (e.g. for recharge control). The model strikes a good balance between generality, realism and accuracy, and provides a powerful tool for recharge study.

In what follows, we show how the WAVES model was used to investigate the effects of management on the water balance of irrigated crops on the NCP and to evaluate deep drainage under different cropping rotations in the Australian Mallee region.

Example 1: Modelling the water balance of irrigated crops on the North China Plain

Wang et al. (2001) used the WAVES model to analyse data obtained from field experiments at Luancheng Eco-Agro-System Experimental Station on the NCP and also to simulate the effect of irrigation management on crop growth in the region. Most rain falls in summer so irrigation is required during the winter wheat growing season, when the difference between rainfall and evapotranspiration is large. Corn grows during summer, but some irrigation is still required.

A relevant issue in irrigated agriculture is the relative importance of soil evaporation and transpiration. Results obtained from WAVES modelling suggest that before canopy closure — where the leaf area index (LAI) is less than 1 — soil evaporation accounted for 50–90% of total evapotranspiration. There appears to be some scope to improve irrigation efficiency by altering the balance between these two fluxes. One strategy is to reduce soil evaporation by mulching. Covering the surface with plant residues can reduce radiation and wind at the surface and hence reduce soil evaporation. Reduction of soil evaporation during the first stage when the soil is wet and evaporation is controlled by atmospheric demand can provide the crops with an opportunity to use the moisture in the top soil layers. During the second stage, when the soil is dry and evaporation is controlled by the moisture content, the rate of evaporation is usually much lower than during the first stage and the effect of mulching is likely to be small. The field experiments showed that mulching reduced soil evaporation by 50% under winter wheat (Fig. 6); this is equivalent to 80 mm of water over the entire growing season. In terms of irrigation efficiency, this means that we can reduce irrigation water by 25% over the entire growing season.

To further investigate the effect of irrigation on crop yield (growth), we conducted several scenario simulations using WAVES. The amount of water applied in each irrigation varied from 0–80 mm; Figure 7 shows the resulting leaf area development. It should be noted that irrigation had no impact on crop growth in wet years; in dry years it enhanced crop growth significantly, but the benefit became less obvious as irrigation water supply increased. The results suggest that current irrigation practices in the area tend to overirrigate crops. Given the limited water available for irrigation in the region, resulting in falling groundwater levels, irrigation cannot be maintained sustainably at current levels.

Example 2: Simulating episodic recharge under different crop rotations

Zhang et al. (1999a,b) described two field experiments conducted at Hillston (New South Wales) and Walpeup (Victoria) to see whether changing land use and agronomic practices could reduce groundwater recharge. Various crop and pasture rotations involving fallow, field pea, Indian mustard, wheat, oats, lucerne and medic pastures were considered. The WAVES model was calibrated with the field data and then used to simulate soil moisture content, plant growth and recharge under these rotations.

The WAVES model was able to accurately simulate soil moisture contents at both sites throughout the study period (Fig. 8). The depths were chosen to represent different soil layers and root zones at the two sites. The model was able to reproduce seasonal variations in soil moisture for different soil types under various cropping rotations. To further evaluate the performance of the model in simulating soil water dynamics, we compared calculated and

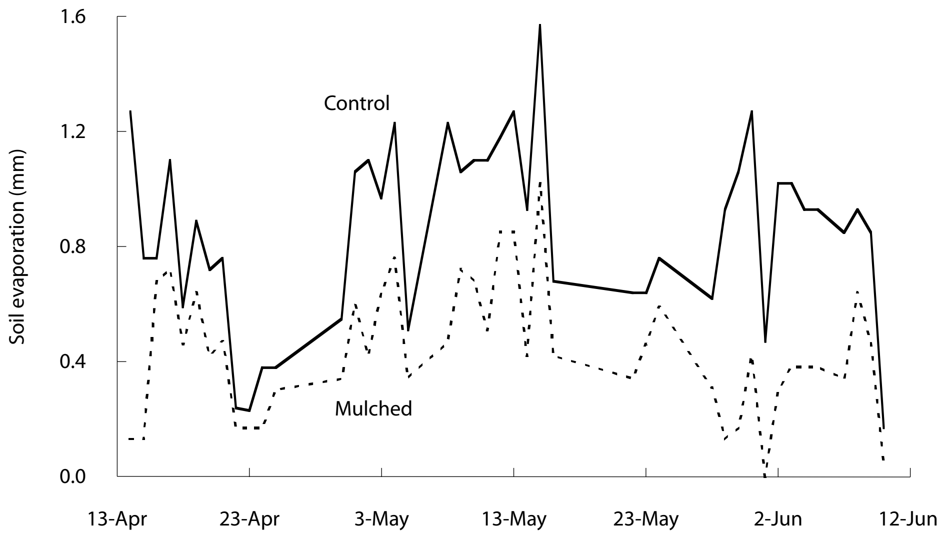

Figure 6. Effect of mulching on daily soil evaporation under winter wheat, 1996.

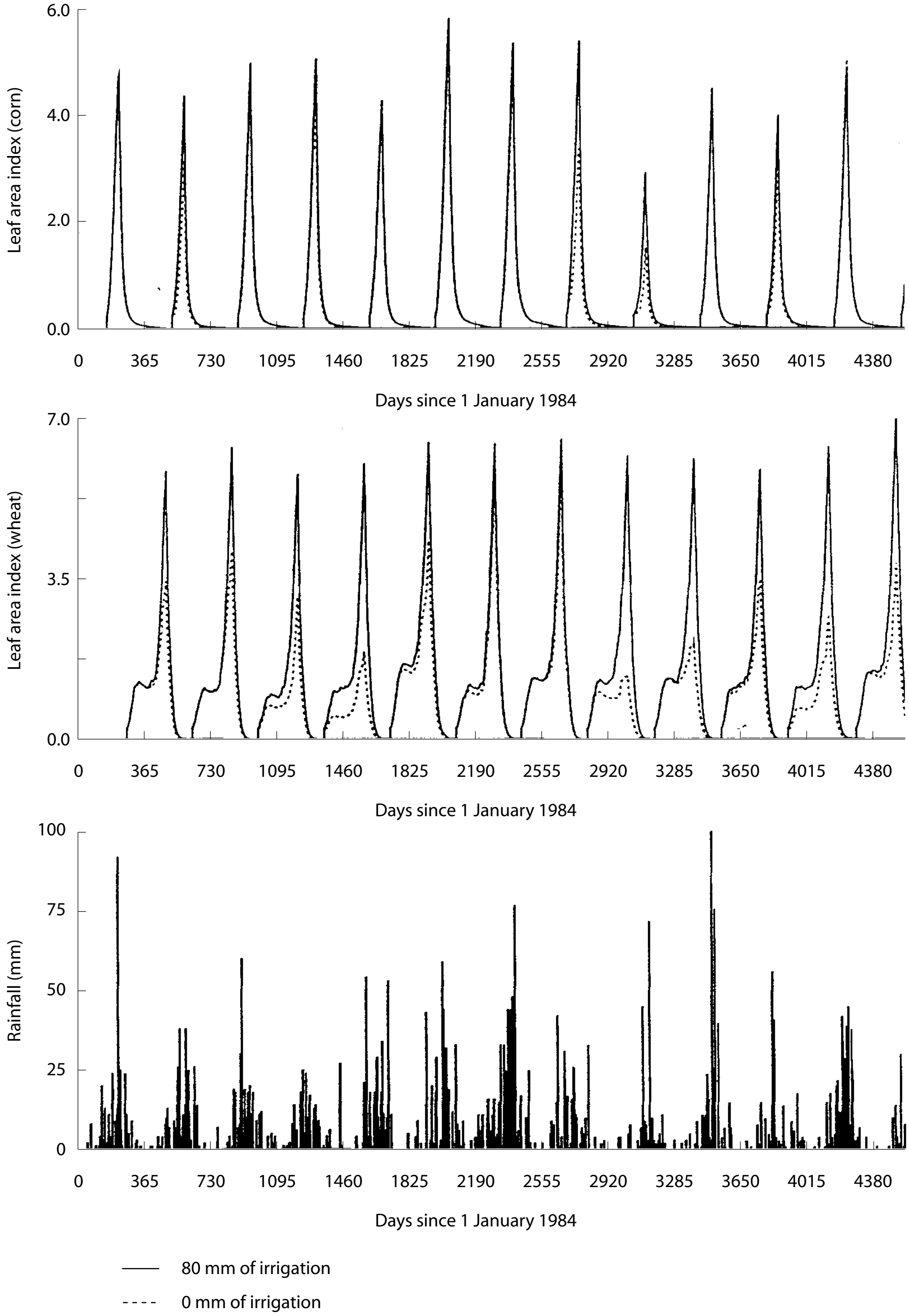

Figure 7. Effects of irrigation on leaf area development as simulated by WAVES.

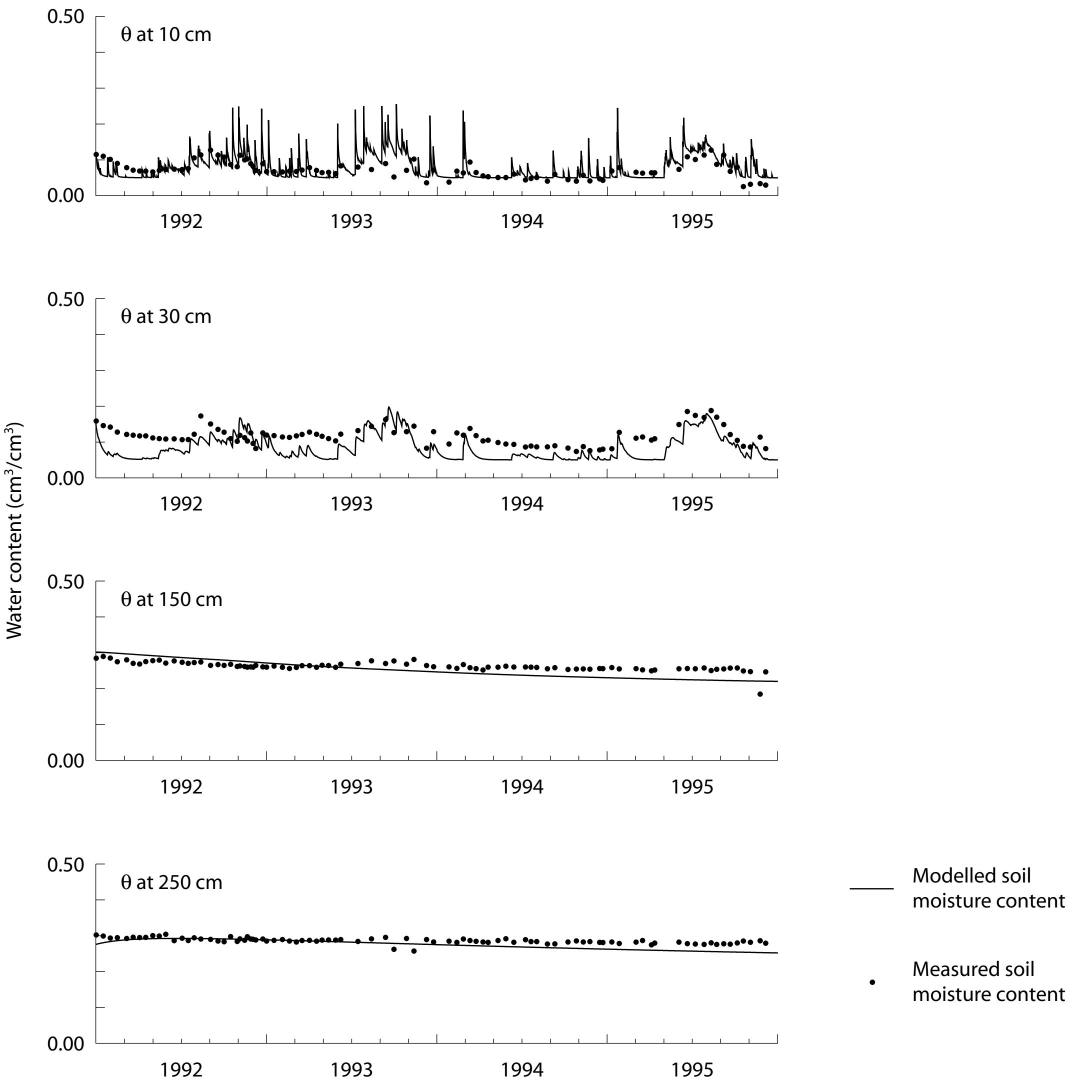

Figure 8. Comparison between modelled and measured soil moisture content at Hillston.

measured soil moisture profiles for different periods. The model agreed very well with the measurements throughout the soil profile (Fig. 9). At Hillston, throughout the study period, there was a drying front associated with maximum rooting density at approximately 1 metre, below which the soil water remained relatively constant.

At Hillston, the simulated recharge rate at 4.0 metre depth increased dramatically after 10 years for the medic rotation but not for the lucerne rotation, which continued to decrease (Fig. 10a). The recharge under medic rotation appeared to respond to the cumulative rainfall anomaly. At Walpeup, a similar trend was observed for the fallow rotation with shallow rooting depth. However, an increase in simulated recharge occurred after 20 years with deep-rooted plants (Fig. 10c). The nonfallow rotation was not sensitive to the cumulative rainfall anomaly, but the fallow rotation was (Figs 10b and 10c). The difference in simulated recharge under fallow and nonfallow rotations is significant.

These results suggest that changes in agronomic practice (for example, fallowing and crop rotation) may take a considerable period of time (more than

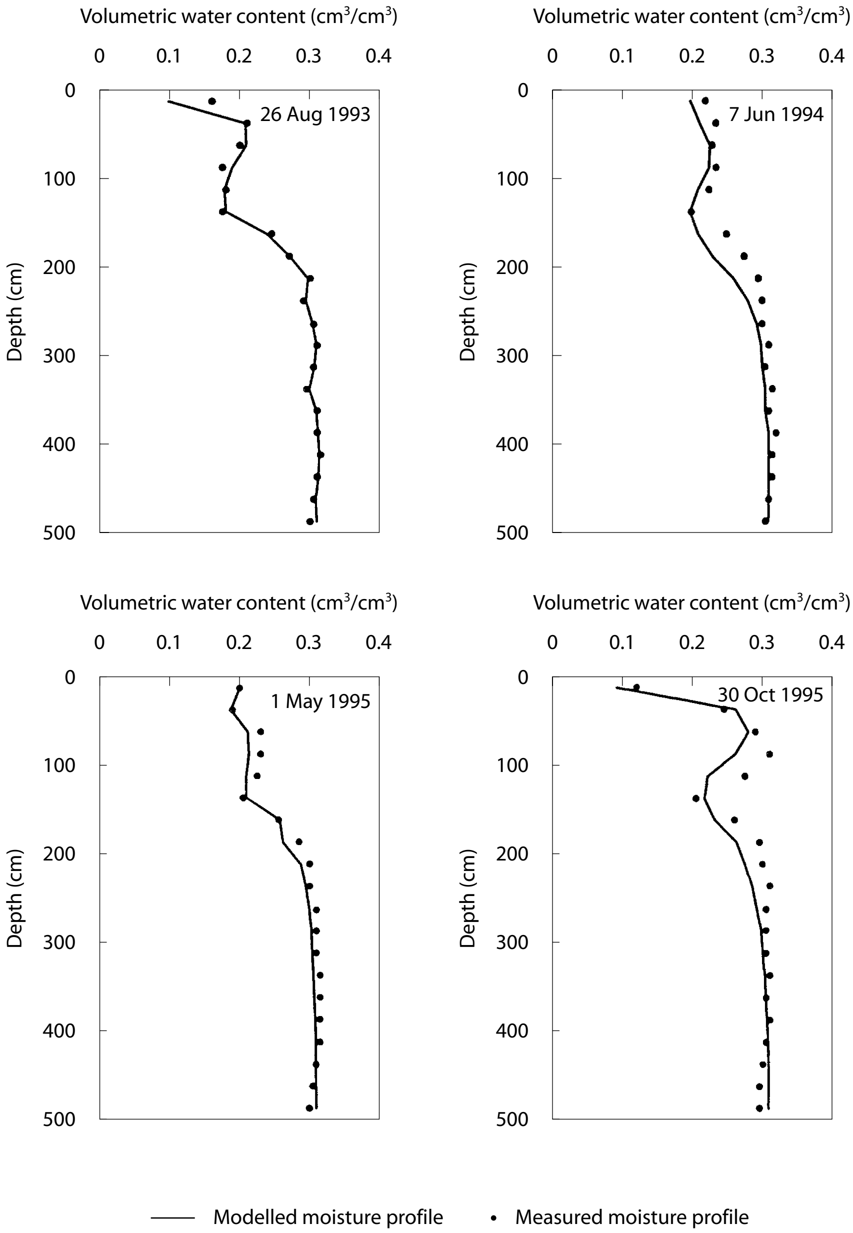

Figure 9. Modelled and measured soil moisture profiles at Walpeup for the selected dates.

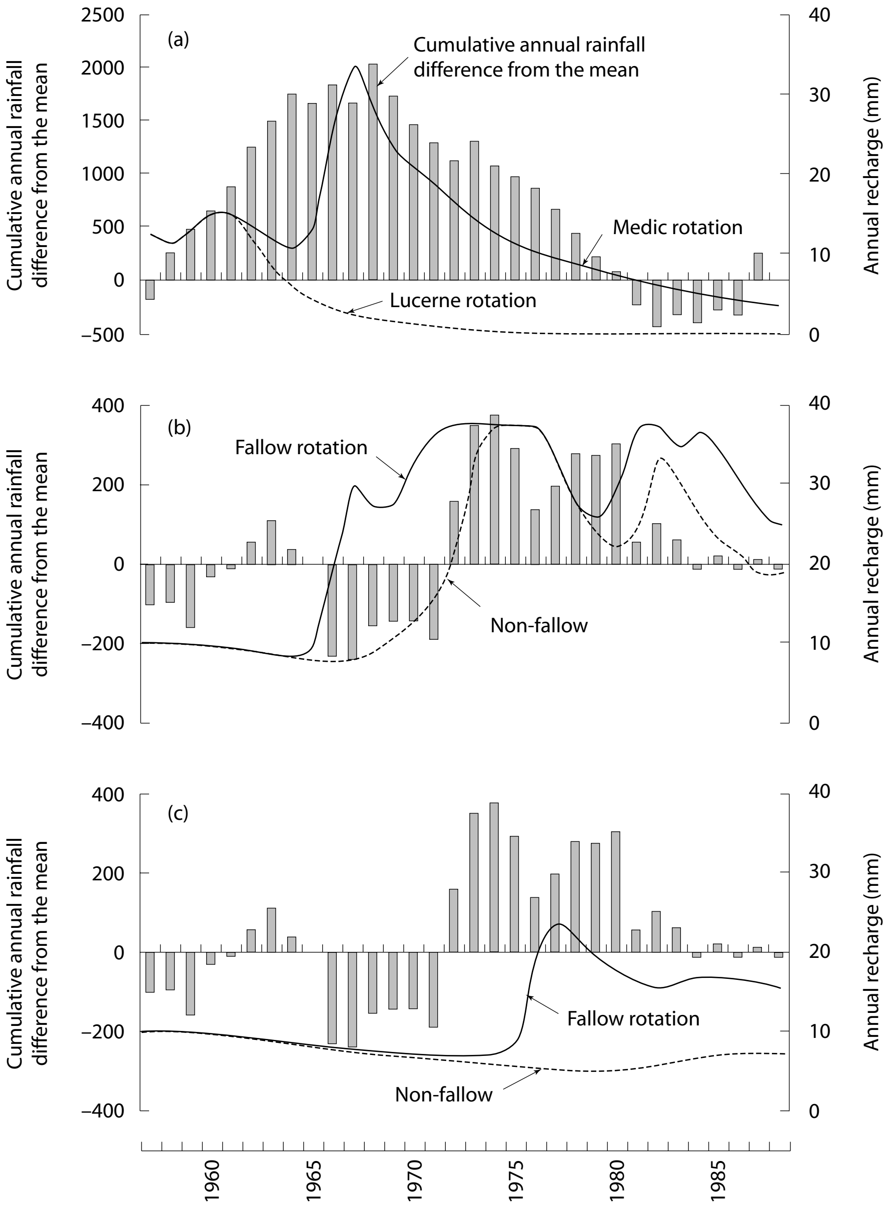

Figure 10. Cumulative annual rainfall differences from the mean and annual recharge rates at 400 cm depth for (a) Hillston under lucerne rotation (- - -) and medic rotation (—); (b) Walpeup with a rooting depth of 50 cm under nonfallow (- - -) and fallow rotation (—); and (c) Walpeup with a rooting depth of 100 cm under nonfallow (- - -) and fallow rotation (—).

10 years) to have any noticeable impact on recharge. The results also showed that deep-rooted plants have better control of recharge, but that the degree of control is modified by soil characteristics and the prevailing weather conditions.

The results showed that the recharge just below the root zone is episodic: it occurs infrequently and its magnitude is significant. Given the fact that plants can only use water in the root zone, the effect of current agronomic practices on episodic recharge is limited. During large rainfall events, the root zone (generally considered as a buffer zone) became saturated and significant recharge occurred. Episodic recharge can therefore substantially reduce the effectiveness of land management options in controlling recharge. This is more so for sandy soils than for clay soils because of low water holding capacity and high infiltration rates.

Acknowledgments

This project was supported by ACIAR (LWR1/95/07).

References

Beven, K. 1981. Kinematic subsurface stormflow. Water Resources Research, 17, 1419–1424.

Broadbridge, P. and White, I. 1988. Constant rate rainfall infiltration: a versatile nonlinear model. 1. Analytic solution. Water Resources Research, 24, 145–154.

Dawes, W.R. and Short, D.L. 1993. The efficient numerical solution of differential equations for coupled water and solute dynamics: the WAVES model. Canberra, CSIRO Division of Water Resources, Technical Memorandum 93/18,.

Dunne, T. 1978. Field studies of hillslope flow processes. In: Kirkby, M.J., ed., Hillslope Hydrology. Chichester, Wiley, 227–294.

Dunne, T. 1983. Relation of field studies and modeling in the prediction of storm runoff. Journal of Hydrology, 65, 25–48.

Freeze, R.A. 1974. Streamflow generation. Reviews of Geophysics and Space Physics, 12, 627–647.

Hook, R.A., Fleming, P.M. and Thomas, C. 1998. Understanding models. In: Willams, J., Hook, R.A. and Gascoigne, H.L., eds, Farming Action—Catchment Reaction: the effect of dryland farming on the natural environment. Melbourne, CSIRO Publishing.

Horton, R.E. 1919. Rainfall interception. Monthly Weather Review, 47, 603–623.

Jarvis, P.G. and McNaughton, K.G. 1986. Stomatal control of transpiration: scaling up from leaf to region. Advances in Ecological Research 15, 1–49.

Lerner, D.N., Issar, A.S. and Simmers, I. 1990. Groundwater recharge—a guide to understanding and estimating natural recharge. Hannover, International Association of Hydrogeologists, Heise.

Monteith, J.L. 1981. Evaporation and surface temperature. Quarterly Journal of the Royal Meteorological Society, 107, 1–27.

Richards, L.A. 1931. Capillary conduction of liquids through porous media, Physics, 1, 318–333.

Rowell, D.L. 1994. Soil Science: methods and applications. New York, John Wiley & Sons, Inc.

Scotter, D.R., Clothier, B.E. and Turner, M.A. 1979. The soil water balance in a fragiaqualf and its effect on pasture growth in Central New Zealand. Australian Journal of Soil Research, 17, 455–465.

Short, D.L., Dawes, W.R. and White, I. 1995. The practicability of using Richards' equation for general purpose soil-water dynamics models. Environment International, 21, 723–730.

Sophocleous, M.A. 1991. Combining the soilwater balance and water-level fluctuation methods to estimate natural groundwater recharge: practical aspects. Journal of Hydrology, 124, 229–241.

Walker, G.R. and Zhang, L. 2001. Plot-scale models and their application to recharge studies. In: Zhang, L. and Walker, G.R., eds, Studies in Catchment Hydrology: the basics of recharge and discharge. Melbourne, CSIRO Publishing.

Wang, H.X., Zhang, L., Dawes, W.R. and Liu, C.M. 2001. Improving water use efficiency of irrigated crops in the North China Plain—measurements and modelling. Agricultural Water Management, 48, 151–167.

White, R.E., Helyar, K.R., Ridley, A.M., Chen, D., Heng, L.K., Evans, J., Fisher, R., Hirth, J.R., Mele, P.M., Morrison, G.R., Cresswell, H.P., Paydar, Z., Dunin, F.X., Dove, H. and Simpson, R.J. 2000. Soil factors affecting the sustainability and productivity of perennial and annual pastures in the high rainfall zone of south-eastern Australia. Australian Journal of Experimental Agriculture, 40, 267–283.

Wu, H., Rykiel Jr., E.J., Hatton, T. and Walker, J. 1994. An integrated rate methodology (IRM) for multi-factor growth rate modelling. Ecological Modelling, 73, 97–116.

Zhang, L., Dawes, W.R. and Hatton, T.J. 1996. Modelling hydrological processes using a biophysically based model—application of WAVES to FIFE and HAPEX-MOBILHY. Journal of Hydrology, 185, 147–169.

Zhang, L., Dawes, W.R., Hatton, T.J., Hume, I.H., O'Connell, M.G., Mitchell, D.C., Milthorpe, P.L. and Yee, M. 1999a. Estimating episodic recharge under different crop/pasture rotations in the Mallee region. 2. Recharge control by agronomic practices. Agricultural Water Management, 42, 237–249.

Zhang, L., Hume, I.H., O'Connell, M.G., Mitchell, D.C., Milthorpe, P.L., Yee, M., Dawes, W.R. and Hatton, T.J. 1999b. Estimating episodic recharge under different crop/pasture rotations in the Mallee region. 1. Experiments and model calibration. Agricultural Water Management, 42, 219–235.