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A conceptual model of daily water balance following partial clearing from forest to pasture

M. A. Bari[1,2] **and K. R. J. Smettem**[3,4]

[1]Department of Environment, 3 Plain Street, East Perth, W.A. 6004, Australia
[2]School of Earth and Geographical Sciences, Hydroscience Discipline Group, The University of Western Australia, 35 Stirling Highway, Crawley, W.A. 6009, Australia
[3]School of Environmental Systems Engineering, The University of Western Australia, 35 Stirling Highway, Crawley, W.A. 6009, Australia
[4]Cooperative Research Centre for Plant-Based Management of Dryland Salinity, The University of Western Australia, 35 Stirling Highway, Crawley, W.A. 6009, Australia

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Abstract. A simple conceptual water balance model representing the streamflow generation processes on a daily time step following land use change is presented. The model consists of five stores: (i) Dry, Wet and Subsurface Stores for vertical and lateral water flow, (ii) a transient Stream zone Store (iii) a saturated Goundwater Store. The soil moisture balance in the top soil Dry and Wet Stores are the most important components of the model and characterize the dynamically varying saturated areas responsible for surface runoff, interflow and deep percolation. The Subsurface Store describes the unsaturated soil moisture balance, extraction of percolated water by vegetation and groundwater recharge. The Groundwater Store controls the baseflow to stream (if any) and the groundwater contribution to the stream zone saturated areas. The daily model was developed following a *downward approach* by analysing data from Ernies (control) and Lemon (53% cleared) catchments in Western Australia and elaborating a monthly model. The daily model performed very well in simulating daily flow generation processes for both catchments. Most of the model parameters were incorporated a priori from catchment attributes such as surface slope, soil depth, porosity, stream length and initial groundwater depth, and some were calibrated by matching the observed and predicted hydrographs. The predicted groundwater depth, and streamflow volumes across all time steps from daily to monthly to annual were in close agreement with observations for both catchments.

Correspondence to: M. A. Bari
(mohammed.bari@environment.wa.gov.au)

1 Introduction

Over the last three decades considerable research had been undertaken in Western Australia to understand changes in streamflow and salinity generation processes following agricultural clearing. Most of the research was devoted to establishment and intense monitoring of a number of experimental catchments with different land use options. It is now well understood that forest clearing for agriculture has led to an increase in groundwater recharge and rising water tables. This process mobilises the salt stored in the unsaturated zone of the soil profile and eventually discharged it to streams (Wood, 1924; Peck and Williamson, 1987; Schofield and Ruprecht, 1989; Ruprecht and Schofield, 1991; Bari, 1998). The magnitude of stream salinity increase is dependent on annual rainfall and the extent and location of clearing (Schofield and Ruprecht, 1989).

Different hydrological models have also been developed in the past to represent the changes in physical processes associated with different land use and climate changes. Most of the early models were lumped and statistical. A distributed conceptual model, the Darling Range Catchment Model (DRCM), was developed and applied to some catchments in the Darling Range of Western Australia (Mauger, 1986). Sivapalan et al. (1996) simplified the conceptual form of DRCM and developed the Large Scale Catchment Model (LASCAM). This model was tested, calibrated and validated across a range of different catchments, from small experimental to very large (Sivapalan et al., 2002). Topog (Vertessy et al., 1993) and WEC-C (Water and Environmental Consultants-Catchment) are two other fully distributed models which are applicable to hill slope and experimental scale (Croton and Barry, 2001; Croton and Bari, 2001).

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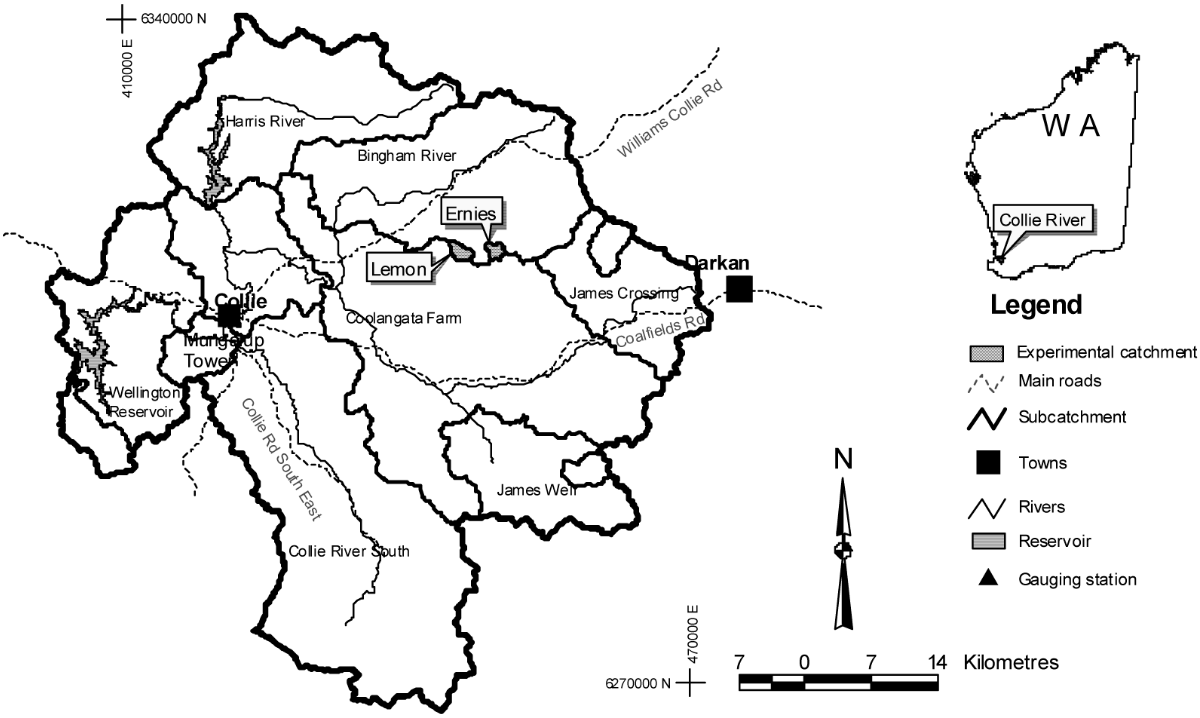

Fig. 1. Location of the experimental catchments.

Although distributed hydrological models are applied all over the world, it is now well understood that the basic limitations of these models to represent catchment response with a small number of parameters, is due to their inability to reproduce dynamic variation of saturated areas within the catchment (Beven, 1989; Binley et al., 1989; Beven, 2001). In fact, the dynamic variation of the saturated area, a function of accumulation and horizontal movement of water in the top soil layers, is mainly responsible for the highly non-linear nature of catchment response to storm events (Ruprecht and Schofield, 1989; Todini, 1996). Most of the existing conceptual and semi-distributed models require a large number of parameters to represent dynamic variation of the saturated areas. Many of these parameters lack physical meaning as they represent averages at catchment or subcatchment scale. Although different automatic calibration techniques have been developed to estimate model parameter sets of particular applications (Duan, 2003), recent comparisons of the performance of different conceptual rainfall runoff models reveal that model performance depends more on model structure and data quality than on model complexity (Perrin et al., 2001; Gan and Biftu, 2003).

The *downward approach* in model building, originally adopted by Klemes (1983), has revealed new insights into the parsimony of conceptual model structures in Western Australia and other parts of the world (Jothityangkoon et al., 2001; Atkinson et al., 2002; Farmer et al., 2003). The model building procedure shows that scale of interest, both time (annual to hourly) and space (point to ~1000 km^2), determines the model complexity requirements. These recent works have been devoted to water balance prediction of steady-state catchments only.

Data collected from experimental catchments in the southwest of Western Australia show different rates of groundwater level rise, originally not connected to stream invert, following clearing of deep-rooted native forest for pasture development. When the rising groundwater level reaches the stream invert and creates groundwater-induced saturated areas, streamflow and salt discharge increases greatly (Croton and Bari, 2001; Ruprecht and Schofield, 1989). Bari and Smettem (2004) followed the *downward approach* to identify the minimal model structure and complexity required to represent the changes in streamflow generation process following land use changes on a monthly time step.

In this paper we extend the work of Bari and Smettem (2004) to examine the additional complexity and structural changes required, then develop a model to represent runoff following land use change on a daily time step. Data obtained from two experimental catchments (Lemon and Ernies – treated and control respectively) were analysed further to understand the processes and then the monthly model was elaborated accordingly. The daily model consists of three main components: (i) Dry, Wet and Subsurface Stores for vertical and lateral unsaturated water flow, (ii) Stream zone Store and (iii) a saturated Groundwater Store. The main inclusion is the Dry and Wet Stores and a probability distribution function

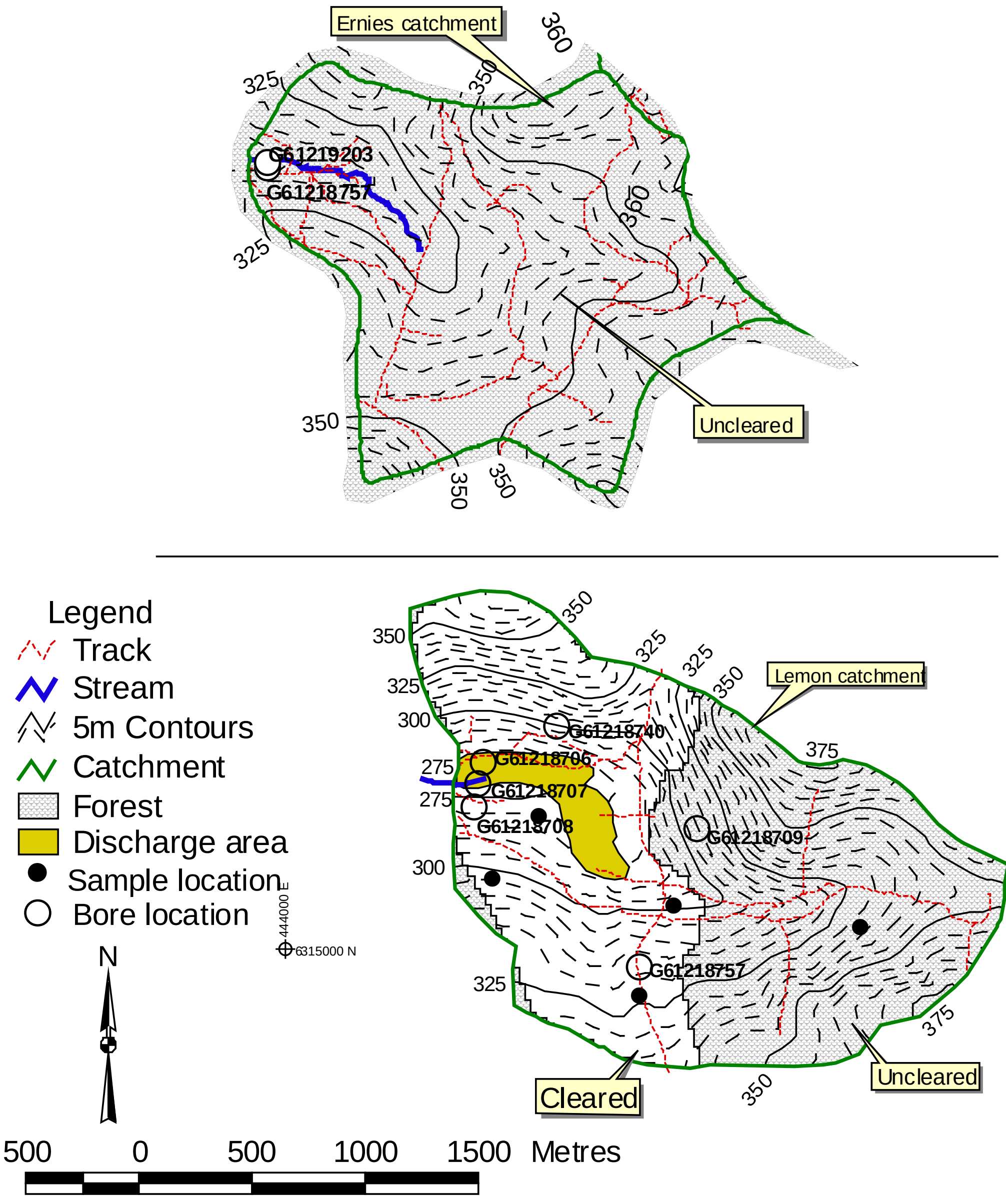

Fig. 2. Detail set up of Ernies and Lemon catchments.

for catchment soil moisture stores and dynamic variation of the conceptual groundwater level. The objective of the daily model is to represent streamflow changes following clearing of native forest with the minimal number of parameters necessary to represent the key processes whilst retaining some physical meaning. Some of the parameters can be estimated a priori but others do require some calibration. We are ultimately interested developing a basin-scale model in which this model will serve as "building block" and additional parsimony can be achieved by giving regionalized estimates for some parameters.

2 Catchment description

The Lemon and Ernies catchments are located in the Collie River catchment, south-west of Western Australia, about 250 km south of Perth (Fig. 1). These catchments have a Mediterranean climate, with cool, wet winters and warm to hot, dry summers. The Class A annual pan evaporation and annual rainfall are approximately 1600 mm (Luke et al., 1988) and 650 mm respectively. The soil profile typically consists of 0.5–6.5 m highly permeable top soil overlying 10–30 m of clay with low permeability. The areas of Lemon and Ernies catchments are 344 ha and 270 ha respectively, and surface slopes average 12% and 5% respectively. The

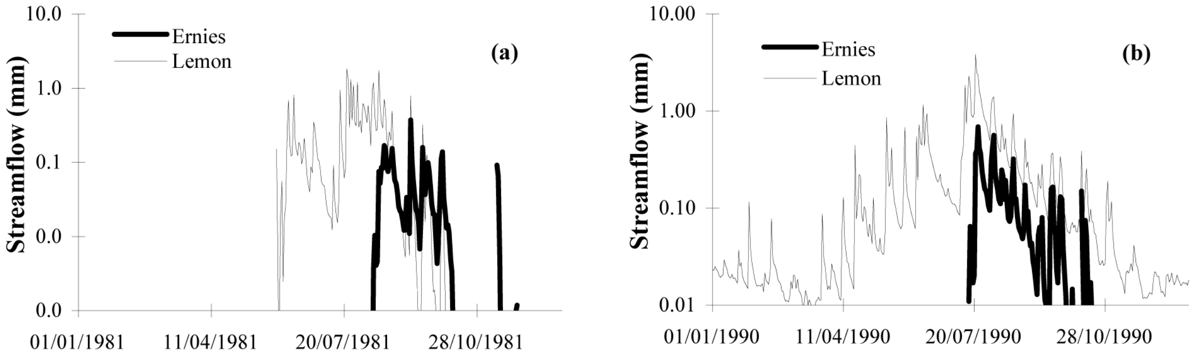

Fig. 3. Comparison of daily streamflow between Ernies and Lemon catchments: **(a)** 1981, **(b)** 1990.

vegetation was an open forest dominated by jarrah (*Eucalyptus marginata*). In the summer of 1976–1977, 53% of the Lemon catchment was logged and was sown to clover and grasses for grazing sheep (Fig. 2). The objective was to understand changes on flow and salinity generation processes following clearing. The Ernies catchment remained as a forested control. Both catchments were instrumented to measure salt and water balances. Rainfall and salt fall were recorded by pluviometers and the streamflow was measured at a sharp-crested V-notch weir for the period 1974–1998. Electrical conductivity of the stream water was continuously measured using a toroidal cell. At Lemon catchment, a network of 72 bores was installed to monitor the groundwater levels. Groundwater level and salinity were recorded from some of the bores during 1977–1998. Five sites were selected in each of the catchments to collect soil moisture content and salt storage (Fig. 2). Hydraulic conductivity of the pallid zone and top soils were also measured (Peck and Williamson, 1987).

3 Streamflow generation process

During the pre-treatment period, a similar runoff response was observed for both catchments and the groundwater level was about 15–20 m below the stream invert (Ruprecht and Schofield, 1991). Following clearing, the deep, permanent groundwater system beneath the Lemon catchment started to rise due to lower evapotranspiration. Groundwater level intersected the surface by 1987 and by 1996 achieved a new stability (Bari and Smettem, 2004). After heavy rainfall, a shallow intermittent groundwater system develops on cap rock or clay, saturates part of the stream zone and generates streamflow by saturation excess overland flow and interflow processes. Immediately following clearing, the flow duration of Lemon catchment increased and started flowing about a month earlier than Ernies catchment (Fig. 3a). The groundwater induced stream zone saturated area increased from nil before clearing to 8% of the catchment area in 1990s. There was an approximately 1400 mm increase in unsaturated soil water storage (Bari and Smettem, 2004). When the groundwater system reached the stream bed, streamflow increased further, became perennial and in the dry summer months was dominated by the baseflow (Fig. 3b).

4 Model description

A *downward approach* originally advocated by Klemes (1983) was followed in developing the daily water balance model. Annual data from experimental catchments (with different land use) were analysed and a simple water balance model was developed which needed minimal calibration (Bari et al., 2005). Further analyses of monthly data demonstrated that a minimal model structure of four interconnecting stores was necessary to represent the landscape hydrological processes (Bari and Smettem, 2004). The four stores for the monthly water balance model were: (i) Upper Store, (ii) Subsurface Store, (iii) Groundwater Store and (iv) Stream zone Store (Fig. 4). The Upper Store generates surface runoff (Q_{r1}), interflow (Q_i) and percolation to the Subsurface Store (I). Trees use most of the percolated water and little recharges the Groundwater Store. When groundwater discharges to the stream, a transient Stream zone Store is created. Additional surface runoff (Q_{r2}) is generated from the "impervious" stream zone saturated area.

We applied the monthly model with the calibrated parameter set for both Ernies and Lemon catchments on a daily time step. The model predicted flow duration quite well but was unable to reproduce the daily peakflow and recessions (Fig. 5). Therefore, we introduced additional complexity and structural changes into the model to better represent daily processes. The Upper Store was partitioned into Dry and Wet Stores, recharge to groundwater was also divided into preferential and matrix flow, and the interception component was elaborated (Fig. 6).

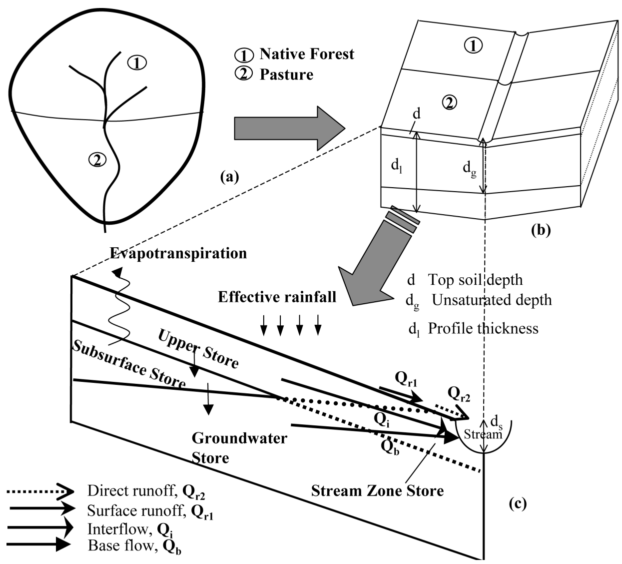

Fig. 4. Schematic representation of a hill slope by four-store model (after Bari and Smettem, 2004).

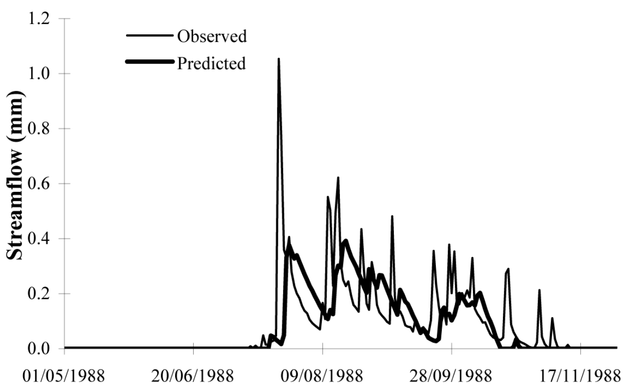

Fig. 5. Observed and predicted daily streamflow at Ernies catchments.

4.1 Evapotranspiration

Evapotranspiration is a major component of the hydrological cycle in the south-west of Western Australia. About 90% of the annual rainfall is lost by evapotranspiration (Sharma, 1983). Annual average interception by mature jarrah forest ranges from 13% to 15% of annual rainfall (Croton and Norton, 1998). In the monthly model we set interception to 13% of rainfall (Bari and Smettem, 2004). For the daily model additional complexity was added as a function of daily rainfall, Leaf Area Index (*LAI*) and interception storage of the forest canopy. The maximum interception or canopy storage capacity (C_{smx}) is determined by assuming that canopy saturation occurs once a certain amount of water accumulates over the plant foliage surface:

$$C_{smx} = C_i LAI \quad (1)$$

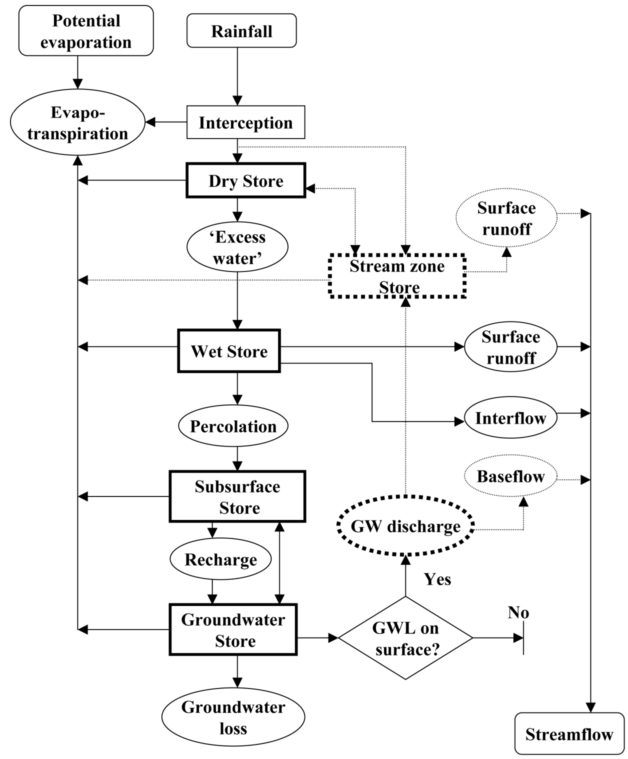

Fig. 6. Flow chart of the hydrological sub-processes in the water balance model.

The actual interception (I_a) is modelled by a simple accounting procedure. Actual canopy storage, $C_s(t, t+1)$, during the period $(t, t+1)$, depends upon the rainfall and actual storage of the previous time step:

$$C_s(t+1) = C_s(t) + R(t, t+1) \text{ if } [C_s(t) + R(t)] < C_{smx} \quad (2a)$$

$$C_s(t+1) = C_{smx} \quad \text{if} \quad [C_s(t) + R(t)] > C_{smx} \quad (2b)$$

$$I_a = C_s - PET \quad \text{if} \quad PET < C_s \quad (3a)$$

$$I_a = C_s \quad \text{if} \quad PET > C_s \quad (3b)$$

Effective rainfall (*RE*) passes through the plant canopy and becomes available for infiltration. Evapotranspiration demand is reduced and residual potential evapotranspiration (*RET*) is the energy available for plant transpiration and soil evaporation.

$$RE(t+1) = R(t, t+1) - I_a(t+1) \quad (4)$$

$$RET = PET - I_a \quad (5)$$

Representation of the actual soil evaporation and plant transpiration of the daily model remained identical to the monthly model (Bari and Smettem, 2004). These two processes take place from all five stores of the model. The soil evaporation depends upon the moisture content of the stores, and Leaf Area Index. Plant transpiration is a function of relative root volume in different stores (RT_u, RT_t), residual potential evaporation, soil moisture content of different stores, Leaf Area Index and a biological parameter (α_t). For example soil evaporation from Dry (E_{sd}) and transpiration from Wet Stores (E_{tw}) are expressed as:

$$E_{sd} = RET \frac{W_d}{W_{dmx} + W_{wmx}} e^{-csLAI} \tag{6a}$$

$$E_{tw} = \alpha_t RET \frac{RT_u}{RT_t} \left[1 - \left(1 - \frac{W_w}{W_{wmx}} \right)^{tu} \right] \frac{LAI}{LAI_{mx}} \tag{6b}$$

4.2 Unsaturated soil water accounting

The unsaturated profile plays an important role in streamflow generation processes in Western Australia. Depth of the profile varies across the different rainfall zones. In the High Rainfall Zone (>1100 mm yr^{-1}) the permanent groundwater level lies within 2 m of the stream invert. Therefore, the vertical thickness of the unsaturated profile is the shortest. In the Low Rainfall Zone (<900 mm yr^{-1}), under pristine land use, the thickness of the unsaturated profile is in excess of 20 m (Bari and Smettem, 2004). Soil profile data analyses reveal the presence of two distinctive soil horizons. The top soil consists of 1–7 m thick highly conductive gravelly and sandy laterite. This layer overlies less permeable sandy loams and kaolinitic clay (Johnston, 1987). Often a cemented layer of aluminosilicate (hard pans) exists below 2–3 m of the top soil, particularly along the stream zone (Pettit and Froend, 1992). Therefore, in the monthly model we divided the unsaturated soil into two stores: (i) Upper Store (ii) Subsurface Store (Fig. 4).

When the monthly model was applied on a daily time step with updated parameter sets, the peakflow and recessions could not be predicted (Fig. 5). We postulated that the sharp hydrograph rise to peaks and similarly sharp, early recessions were due to the formation of dynamic saturated areas along the stream zone following significant rainfall. At the treated (Lemon) catchment, further complexity in the daily flow generation process was evident when the groundwater level reached the streambed. Therefore, a non-linear probability distribution of the depth of the top soil and its water holding capacity was adopted to represent the dynamic variation of the saturated areas. A similar concept has been applied in other models. For example, in the Xinanjiang model the spatial distribution of soil moisture capacity was expressed in two probability distribution functions – one up to the field capacity and the other from field capacity to saturation (Zhao and Liu, 1995). In the VIC and ARNO models a single distribution function was used to describe the soil moisture capacity (Wood et al., 1992; Todini, 1996). The major advantage of this approach is that the catchment soil moisture balance is functionally related by simple analytical expressions to the dynamic contributing areas. Therefore, we incorporate additional complexity and structural changes into the Upper Store of the monthly model to represent the daily soil water movement by two inter-connecting stores: (i) Dry Store and (ii) Wet Store.

4.2.1 Dry Store

We know from field observations that up to the drained upper limit or so-called "field capacity", the soil matrix has the ability to hold water against gravity. The water held against gravity is available for evapotranspiration only. We define this water holding capacity of the soil matrix as the "Dry Store". Soil depth, physical properties and "field capacity" determine the potential volume of the Dry Store. Based on extensive drilling carried out in these experimental catchments, considerable information exists on the depth and distribution of the top soil layer. Typical depth generally ranges from 1 to 7 m and the probability distribution function fits extremely well to the measured soil depth distribution (Sivapalan and Woods, 1995). Due to the very high infiltration capacity of surface soils, we assume that effective rainfall (RE) rapidly infiltrates into the soil matrix. Soil moisture retention capacity (w_d) below any elementary surface area is a function of its field capacity (θ_f) and soil depth (d'), such that:

$$w_d = d'\theta_f \tag{7}$$

Assuming an empirical distribution of soil depths over the catchment, we represent the water holding capacity by a cumulative probability distribution function (Fig. 7a). A catchment of surface area A_t consists of pervious and impervious (A_i) areas. If we denote A_w as the part of the catchment where the water content has reached or exceeded field capacity, then we can represent it as a proportion of catchment area as:

$$x = \frac{(A_w - A_i)}{(A_t - A_i)} = 1 - \left(1 - \frac{w_d}{w_{dm}} \right)^b \tag{8}$$

In the above equation, b is a parameter and w_{dm} is the maximum possible water retention capacity of any elementary area within the catchment. After effective rainfall (RE), part is retained in the Dry Store (Fig. 7) and the other is released (Rf) as:

$$Rf = \frac{A_t - A_i}{A_t} \int_{w_d}^{w_d+RE} x(\sigma) d\sigma \quad \text{if} \quad (w_d + RE) < w_{dm} \tag{9a}$$

$$Rf = \frac{A_t - A_i}{A_t} \left[(w_d + RE - w_{dm}) + \int_{w_d}^{w_{dm}} x(\sigma) d\sigma \right] \quad \text{if} \quad (w_d + RE) > w_{dm} \tag{9b}$$

The above two equations can be expressed in terms of catchment average storage (W_d) and maximum storage

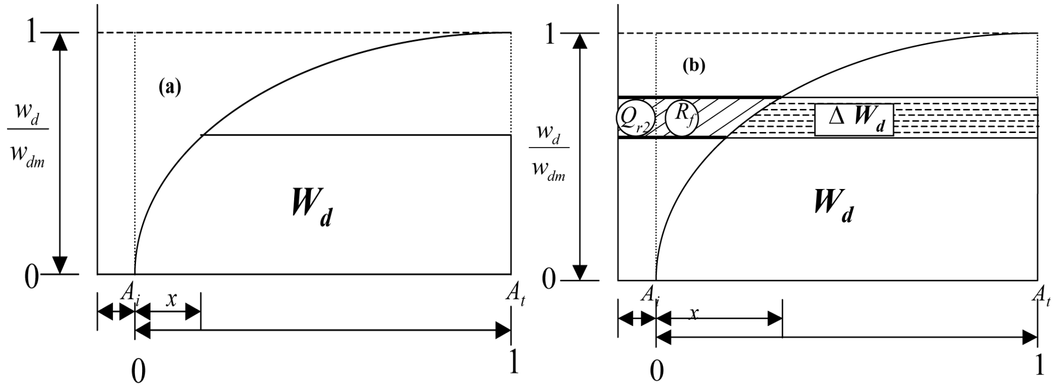

Fig. 7. Cumulative distribution of elementary area **(a)** water retention at field capacity and **(b)** generation of excess water.

(W_{dmx}) in the Dry Store. After integration:

$$Rf = \frac{A_t - A_i}{A_t}\left[RE - W_{dmx} + W_d + W_{dmx}\left\{\left(1 - \frac{W_d}{W_{dmx}}\right)^{\frac{1}{b+1}} - \frac{RE}{(b+1)\,W_{dmx}}\right\}^{b+1}\right]$$

$$\text{if} \quad 0 < RE < (b+1)W_{dmx}\left(1 - \frac{W_d}{W_{dmx}}\right)^{\frac{1}{b+1}} \qquad (10a)$$

$$Rf = \frac{A_t - A_i}{A_t}\left[RE - (W_{dmx} - W_d)\right]$$

$$\text{if} \quad RE \geq (b+1)W_{dmx}\left(1 - \frac{W_d}{W_{dmx}}\right)^{\frac{1}{b+1}} \qquad (10b)$$

Therefore the Dry Store water content update at time (t+1) is:

$$W_d(t+1) = W_d(t) + RE(t,t+1) - E_{td}(t,t+1) - E_{sd}(t,t+1) - Rf(t,t+1) \qquad (11)$$

4.2.2 Wet Store

The Wet Store represents moisture content in the soil matrix from field capacity to saturation. Water held in this store is free to travel through or across the soil matrix. The Wet Store represents the development of an intermittent shallow groundwater table and contributes interflow (lateral flow) to the stream and percolation (vertical flow) to the underlying Subsurface Store. Soil evaporation and transpiration (if any) also take place from this store. The Wet Store controls the formation of the variably contributing dynamic saturated area and surface runoff. It is extended up to the area where the moisture content has reached or exceeded field capacity (Fig. 8). The Wet store occupies a fraction (or whole) of the catchment, part of which is saturated. Like the Dry Store, the capacity of any elementary area where the water content has exceeded field capacity can be written as:

$$w_w = d'(\theta_s - \theta_f) \qquad (12)$$

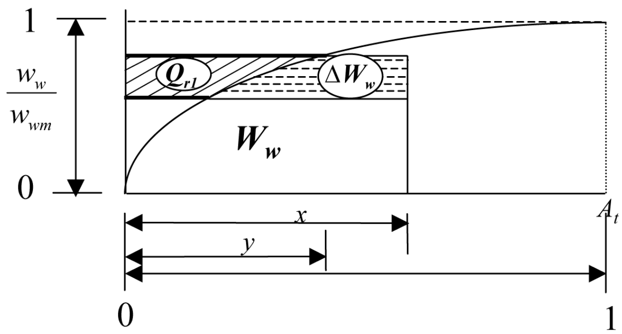

Fig. 8. Generation of surface runoff following a rainfall event.

We assume that the Wet Store capacity is non-uniformly distributed over the area (A_w), where excess water is being produced (Fig. 8). Part of the Wet Store may reach saturation (A_s), the proportion of which can be expressed as:

$$y = \frac{(A_s - A_i)}{(A_w - A_i)} = 1 - \left(1 - \frac{w_w}{w_{wm}}\right)^c, \quad A_s \leq A_w \qquad (13)$$

In the above equation w_w is the elementary area water content at saturation and w_{wm} is the maximum possible water content in any elementary area within the catchment.

The total surface runoff generated by the catchment has two components: (i) from the pervious area (Q_{r1}) and (ii) from the impervious area (Q_{r2}) (Fig. 8). It can be calculated as:

$$Q_r = Q_{r2} + Q_{r1} \qquad (14a)$$

$$Q_r = \frac{A_i}{A_t} RE + x \int_{w_w}^{w_w+Rf} y(\sigma)\,d\sigma$$

$$\text{if} \quad (w_w + Rf) < w_{wm} \qquad (14b)$$

$$Q_r = \frac{A_i}{A_t} RE + x \left[(w_w + Rf - w_{wm}) + \int_{w_w}^{w_{wm}} y(\sigma) d\sigma \right]$$

$$\text{if} \quad (w_w + Rf) > w_{wm} \tag{14c}$$

After integration and transformation we get:

$$Q_r = \frac{A_i}{A_t} RE + \left[1 - \left(1 - \frac{W_d}{W_{dmx}} \right)^{\frac{b}{b+1}} \right] \left[Rf - W_{wmx} + W_w + W_{wmx} \left\{ \left(1 - \frac{W_w}{W_{wmx}} \right)^{\frac{1}{c+1}} - \frac{Rf}{(c+1)\, W_{wmx}} \right\}^{c+1} \right]$$

$$\text{if} \quad 0 < Rf < (c+1) W_{wmx} \left(1 - \frac{W_w}{W_{wmx}} \right)^{\frac{1}{c+1}} \tag{15a}$$

$$Q_r = \frac{A_i}{A_t} RE + \left[1 - \left(1 - \frac{W_w}{W_{wmx}} \right)^{\frac{b}{b+1}} \right] [Rf - (W_{wmx} - W_w)]$$

$$\text{if} \quad Rf \geq (c+1) W_{wmx} \left(1 - \frac{W_w}{W_{wmx}} \right)^{\frac{1}{c+1}} \tag{15b}$$

The representation of daily interflow and percolation (the rate of lateral and vertical drainage from the top soil matrix) remained practically unchanged from the monthly model but we now assume they occur only from the Wet Store and can be expressed as:

$$Q_i = 0 \quad \text{if} \quad W_w < W_{wi} \tag{16a}$$

$$Q_i = K_{ul} \left(\frac{W_w - W_{wi}}{W_{wmx} - W_{wi}} \right)^{ia} x \quad \text{if} \quad W_w > W_{wi} \tag{16b}$$

$$I = K_{uv} \left[1 + pb \left(1 - \frac{W_l}{W_{lmx}} \right)^{pa} \right] \left(\frac{W_w}{W_{wmx}} \right) x \tag{17}$$

Water content update of the Wet Store at time $(t+1)$ is:

$$W_w(t+1) = W_w(t) + Rf(t, t+1) - E_{tw}(t, t+1)$$
$$-E_{sw}(t, t+1) - Q_{r1}(t, t+1) - Q_i(t, t+1) - I(t, t+1) \tag{18}$$

4.2.3 Subsurface Store

The Subsurface Store represents the deep unsaturated soil profile and acts as a delay function for effects of rising groundwater level on streamflow and salinity (Bari and Smettem, 2004). Recharge from the Subsurface to the Groundwater Store occurs in two different processes: (i) from soil matrix as excess flow (Rl_1), and (ii) preferential flow from preferred pathways (Rl_2). Both these processes are accounted for in this model by incorporating additional complexity to the monthly model. Similar to the Dry Store, we define that the soil water capacity of any elementary area is a function of depth ($d'_g - d'$), porosity (ϕ_l) and field capacity (θ_{lf}) and can be described by a distribution function. Therefore, recharge from the soil matrix can be calculated as:

$$Rl_1 = \int_{w_l}^{w_l+I} x'(\sigma) d\sigma \quad \text{if} \quad (w_l + I) < w_{lm} \tag{19a}$$

$$Rl_1 = \left[(w_l + I - w_{lm}) + \int_{w_l}^{w_{lm}} x'(\sigma) d\sigma \right]$$

$$\text{if} \quad (w_l + I) > w_{lm} \tag{19b}$$

After integration and transformation the above two equations become:

$$Rl_1 = \left[I - W_{lmx} + W_l + W_{lmx} \left\{ \left(1 - \frac{W_l}{W_{lmx}} \right)^{\frac{1}{a+1}} - \frac{I}{(a+1)\, W_{lmx}} \right\}^{a+1} \right] \tag{20a}$$

$$Rl_1 = [I - (W_{lmx} - W_l)] \tag{20b}$$

When the water content in the Subsurface Store becomes less than the catchment wide field capacity (W_{ldmx}), the excess water (Rl_1) is recycled for transpiration and recharge to the groundwater store becomes zero. The second component of groundwater recharge represents preferential flow to the Groundwater Store. It is represented by the following formula (Averjanov, 1950):

$$Rl_2 = 0 \quad \text{if} \quad W_l < W_{ldmx} \tag{21a}$$

$$Rl_2 = K_{lv} \left(\frac{W_l - W_{ldmx}}{W_{lmx} - W_{ldmx}} \right)^{3.3} \quad \text{if} \quad W_l > W_{ldmx} \tag{21b}$$

Therefore, total recharge to groundwater store becomes:

$$Rl = Rl_1 + Rl_2 \tag{22}$$

The groundwater level, Subsurface Store and Groundwater Store contents ($\Delta d_g, \Delta W_l, \Delta W_{gl}$) change due to recharge (Rl) to the Groundwater Store, loss of groundwater below the gauging station (Q_{loss}), baseflow to the stream zone (Q_{bl}) and transpiration from groundwater (E_{tg}) (Fig. 6). The representation of these processes in the daily model remained unchanged from the monthly model. The exchange of water between the Subsurface and Groundwater Stores is expressed as:

$$\Delta W_{gl} = \left(\frac{\Delta d_g}{d'_g - d} \right) W_l \quad \text{if} \quad \Delta d_g < 0 \tag{23a}$$

$$\Delta W_{gl} = \left(\frac{\Delta d_g}{d_l - d'_g} \right) W_g \quad \text{if} \quad \Delta d_g > 0 \tag{23b}$$

Therefore the Subsurface Store content at time $(t+1)$ is:

$$W_l(t+1) = W_l(t) + I(t, t+1) - E_{tl}(t, t+1)$$
$$-Rl(t, t+1) + \Delta W_{gl}(t, t+1) \tag{24}$$

Table 1. Goodness of fit for daily streamflow simulations.

Measure of fit	Calibration		Verification		Overall	
	Ernies	Lemon	Ernies	Lemon	Ernies	Lemon
EI	0.58	0.78	1.18	0.95	1.04	0.89
E^2	0.44	0.69	0.22	0.85	0.40	0.76
CC	0.76	0.90	0.60	0.92	0.69	0.88

4.3 Groundwater Store

The initial pre-clearing position of the groundwater store is known and the balance of the store is controlled by discharge loss from the store, recharge and baseflow to the stream zone. The volume of the Groundwater Store depends on the location of the conceptual groundwater level and remained identical to the monthly model (Fig. 4). When the conceptual groundwater level does intersect the stream, it contributes to streamflow and indirectly controls the groundwater-induced saturated areas, predominantly in the stream zone. We also incorporated a groundwater loss function to represent the slow migration of the regional groundwater system and loss through the fractured basement. Baseflow to Stream zone Store and groundwater loss can be expressed as:

$$Q_{bl} = K_{ll} L \left|(d_s - d_g)\right| \tan\beta \quad \text{if} \quad d_g < d_s \tag{25a}$$

$$Q_{bl} = 0 \quad \text{if} \quad d_g > d_s \tag{25b}$$

$$Q_{loss} = C_{loss} W_g \tag{26}$$

Therefore the Groundwater Store update at any time $(t+1)$ is:

$$W_g(t+1) = W_g(t) + Rl(t,t+1) - Q_{loss}(t,t+1) \\ - Q_{bl}(t,t+1) - \Delta W_{gl}(t,t+1) - E_{tg}(t,t+1) \tag{27}$$

4.4 Stream zone Store

This store is transient and covers part of the Dry and Wet Stores. Representation of this store became more complex due to the conceptualization of the Dry and Wet Stores. This store content is also influenced by soil evaporation, and loss/gain to/from the Dry Store due to contraction/expansion of the saturated area. All the effective rainfall (RE) which falls on the stream zone, becomes runoff (Q_{r2}). Stream zone Store water content at any time is expressed by:

$$W_{sg} = 0 \quad \text{if} \quad d_g > d_s \tag{28a}$$

$$W_{sg} = \frac{A_i}{A_t}\left(\frac{w_{dm}}{b+1} + \frac{w_{wm}}{c+1}\right) \quad \text{if} \quad d_g < d_s \tag{28b}$$

Soil evaporation (E_{ss}) and plant transpiration (E_{ts}) also takes place from this store. The residual of the baseflow coming to the stream zone becomes actual baseflow to stream:

$$Q_b(t+1) = Q_{bl}(t,t+1) - E_{ss}(t,t+1) - E_{ts}(t,t+1) \tag{29}$$

We assume a complete 'displacement' of Wet Store and Stream zone Store water contents and free mixing due to contraction or expansion of the saturated area. When the groundwater level increases and the stream zone saturated area expands (ΔA_i), the Dry Store loses water to the stream zone and vice versa. It can be calculated as:

$$\Delta W_{sg} = \left(\frac{\Delta A_i}{A_t}\right)\frac{w_{dm}}{b+1} \quad \text{if} \quad \Delta d_g < 0 \tag{30}$$

Therefore the Stream zone Store water content update at any time (t+1) is:

$$W_{sg}(t+1) = W_{sg}(t) + \Delta W_{sg}(t,t+1) \tag{31}$$

4.5 Total streamflow

Total streamflow is the sum of surface runoff, interflow and baseflow components and can be expressed as:

$$Q_t = Q_r + Q_i + Q_b \tag{32}$$

5 Data requirements and calibration

For the Ernies catchment, the first five years of data were used for calibration. As there were significant changes in land use and flow generation processes at the Lemon catchment, streamflow and groundwater data up to 1987 were used for calibration. The rest of the streamflow data was used for model verification. A trial and error method was adopted for calibration of the model. The model performance was evaluated by matching the: (i) observed and predicted daily streamflow hydrographs and (ii) conceptual and observed groundwater levels. A set of statistical criteria was defined for measuring the agreement between the predicted and observed streamflow for both the Ernies and Lemon catchments. The statistical criteria include: (a) flow-period Error Index (EI), (b) Nash-Sutcliffe Efficiency (E^2) and (c) Correlation Coefficient (CC) (Table 1). For an acceptable calibration, both the E^2 and CC for monthly streamflow were set to be greater than 0.8.

In the model Ernies catchment was represented as one subcatchment (Fig. 4). To represent clearing of forest, Lemon catchment was divided into two subcatchments with 53%

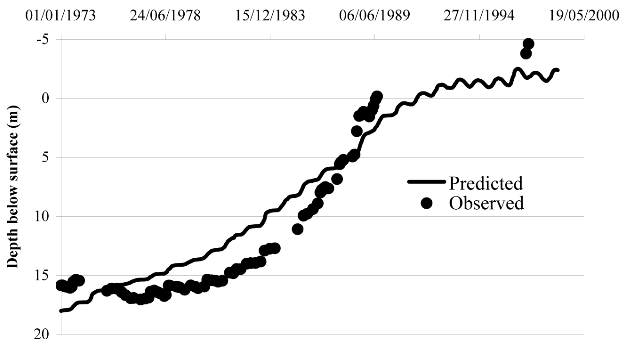

Fig. 9. Observed and predicted groundwater level at the stream zone of Lemon catchment.

and 47% of the area respectively (Fig. 2). Except the rooting depth and Leaf Area Index, all other parameters remained identical for both subcatchments. The rooting depth and LAI of one subcatchment representing clearing were changed from trees to pasture values in 1977 when clearing took place.

Most of the model parameters were estimated a priori from catchment attributes and remained unchanged from the monthly model (Bari and Smettem, 2004). These include groundwater slope (β), stream length (L), porosity (ϕ_l), field capacity (θ_f, θ_{lf}), soil profile thickness (d_l), depth to groundwater level (d_g), land use history, rooting depth-distribution (RT_u, RT_t) and Leaf Area Index (LAI). The groundwater slope that stream lengths were obtained from observed groundwater levels and topographic data. From drilling information, porosity, field capacity, top soil thickness, average depth to the groundwater level and profile thickness were obtained. Mean Leaf Area Index, relative root volumes and rooting depths of pasture and native forests were estimated from available literature. The parameters associated with interception are calibrated against the throughfall measurements undertaken within the jarrah forest of Western Australia (Croton and Norton, 1998).

There are a few parameters in the model whose indicative values can be obtained a priori, but needed calibration by trial and error method for best fit. These include catchment hydraulic properties: (i) lateral (K_{ul}) and vertical (K_{uv}) conductivity of the top soil, and its relationship with moisture content; (ii) lateral conductivity (K_{ll}) of the groundwater system; and (iii) vertical conductivity (K_{lv}) of the deep unsaturated clay profile. The lateral conductivity of the top soil and vertical conductivity of the interface between the top soil and clay profile were calibrated to 395 mm day^{-1} and 27.2 mm day^{-1} respectively for both catchments. The vertical hydraulic conductivity (K_{uv}) falls within the measured value of 0.2–33.7 m day^{-1} (Sharma et al., 1987). The vertical conductivity of the clay layer (K_{lv}) was calibrated to 0.8 mm day^{-1}, slightly less than obtained from slug tests of 2.3–7.6 mm day^{-1} (Peck and Williamson, 1987). One plausible explanation is that the model seeks to represent the catchment average effective conductivity while slug test results represent a collection of point data. The parameter (ia) representing the non-linear relationship between the moisture content and lateral conductivity of the Wet Store was calibrated to 2.15 and 3.15 for the Lemon and Ernies catchments respectively (Eq. 16). The other two parameters (pa, pb) remain unchanged from the monthly model (Eq. 17). Values of the other two important parameters (b, c), which express the degree of homogeneity of soil characteristics over the catchment, were determined by calibration and were very similar for both catchments. Initial soil moisture contents of the unsaturated stores were estimated from soil moisture profile analyses (Bari and Smettem, 2004).

6 Results and discussion

6.1 Groundwater system

The deep groundwater system was about 15–20 m below the surface and was stable before clearing for both catchments. There was some within-year variation, due to groundwater recharge. There was a systematic rise in groundwater levels following clearing at Lemon catchment but the groundwater remained stable beneath native forest at both catchments (Bari and Smettem, 2004). The daily model represented the trend in groundwater level very well in both the catchments. For example, the predicted conceptual groundwater level of Lemon catchment rose faster than observed following clearing, particularly along the streamlines (Fig. 9). The predicted groundwater level reached surface by 1987 creating a transient Stream zone Store and saturated areas. Results from experimental catchments in Western Australia show that the rate of groundwater increase depends on: (i) location and type of clearing and (ii) annual rainfall. The observed rate of groundwater level rise at Lemon catchment (650 mm annual rainfall) was lower than other catchments with higher rainfall and similar clearing (Bari and Ruprecht, 2003; Bari and Smettem, 2004). It also appears to have taken a much longer time (1977–1995) to reach equilibrium mainly due to the low recharge rate, greater soil moisture deficit and greater unsaturated profile thickness.

6.2 Variable contributing saturated area

The groundwater system has two components: shallow and deep ground water systems. The shallow groundwater system is present only in the wet period of the year, when streamflow is generated (Bari and Smettem, 2004). The daily model represented this process very well. For example, a shallow bore located in the lower part of the stream zone of the Ernies catchment retains water only for the wet period of the year. This corresponds to the expansion and contraction of the saturated areas (Fig. 10a). The predicted within-year variation of the stream zone saturated area at Ernies

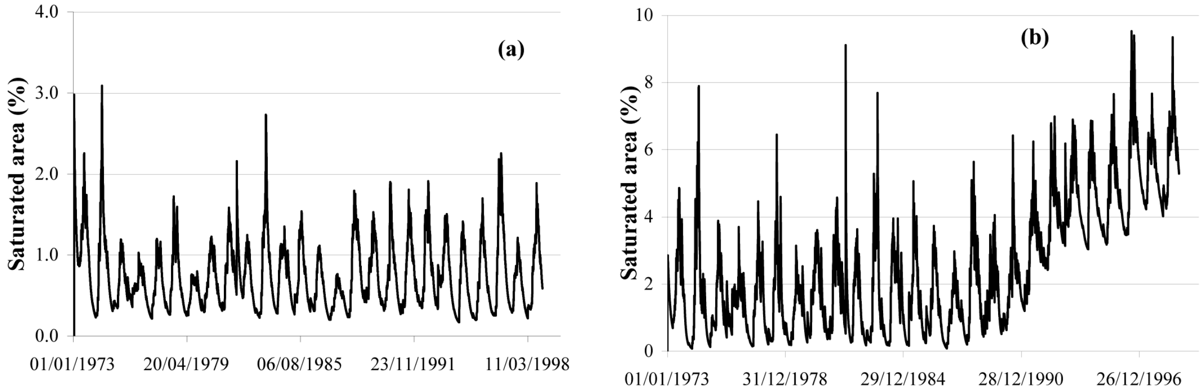

Fig. 10. Predicted variable contributing saturated areas: **(a)** Ernies, **(b)** Lemon catchments.

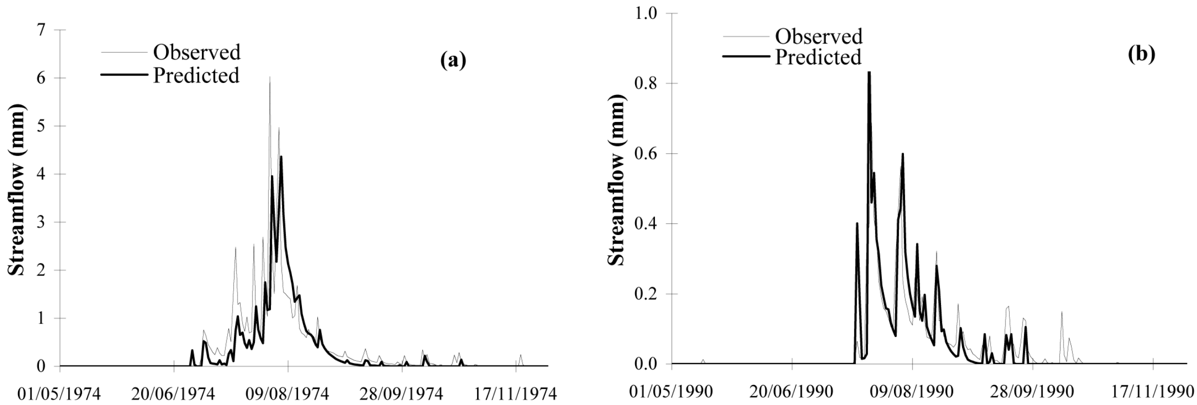

Fig. 11. Actual and simulated daily streamflow – Ernies catchment **(a)** 1974 and **(b)** 1990.

catchment was similar to the monthly model and the annual mean area was estimated at 2% (Bari and Smettem, 2004). Similar trends in variable contributing saturated areas were also observed at Lemon catchment, although the magnitude was generally higher. When the groundwater level reached stream invert, there was a systematic increase in saturated areas in subsequent years (Fig. 10b). When Lemon catchment reached a new stability the daily model predicted the within-year variation of the stream zone saturated area to range from 2–10%. Site visit and interpretation of Landsat photographs indicate that the stream zone of Lemon catchment, that remained saturated over the summer period, increased from 1% in 1991 to 8% by 1998 (Bari et al., 2005). The predicted saturated area matches reasonably well with the estimates from Landsat photographs.

6.3 Streamflow

6.3.1 Daily flow

The Ernies catchment received the highest annual rainfall of 851 mm in 1974. As the permanent groundwater system was far below the stream, streamflow was generated by saturation excess overland flow and interflow processes only. The presence of groundwater in the shallow bore in the stream zone is the evidence of the saturation excess overland flow and interflow generation processes (Bari and Smettem, 2004). The daily model successfully represented the flow generation processes but under predicted the peak flows of the year (Fig. 11a). The observed and simulated hydrographs were very similar for the average-flow year of 1990 (Fig. 11b). The model precisely predicted the timing of the commencement of flow and also the peak flows.

The Lemon catchment received the lowest recorded rainfall in 1979. If not cleared, it may not have produced any runoff at all, as the control catchment did not flow. The model predicted the flow generation process very well, including the flow-duration, peak and recession (Fig. 12a). The daily predicted streamflow was in excellent agreement with the observed values in terms of volume, peak, recession and timing for 1985, when the groundwater level was slightly below the streambed. The catchment received one of the lowest rainfalls of 546 mm in 1997. As the groundwater system rose

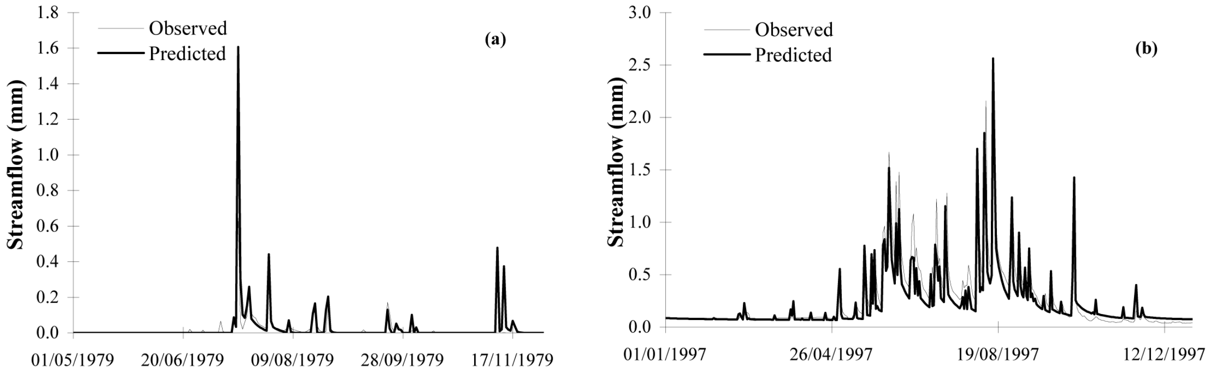

Fig. 12. Actual and simulated daily streamflow – Lemon catchment **(a)** 1979, and **(b)** 1997.

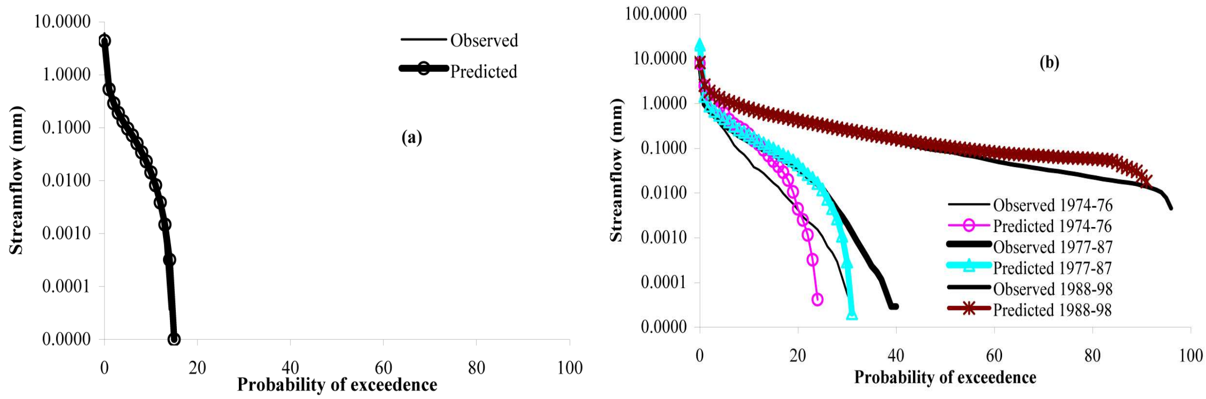

Fig. 13. Observed and predicted flow duration curves – **(a)** Ernies and **(b)** Lemon catchments.

and created permanent groundwater-induced saturated areas (Fig. 9) the streamflow duration increased and ultimately the stream was flowing for the whole year (Fig. 12b). During the period of high-rainfall months (May to October), the model simulated the peak flows well, but over-estimated the interflow component particularly during November and December (Fig. 12b).

In addition to the joint plots of observed and predicted hydrographs, a set of statistical criteria was defined for measuring the agreement between the predicted and observed streamflow for both catchments. The flow-period Error Index provides a measure between the observed and predicted flow periods within a year. For the Ernies catchment the predicted flow-period was 42% lower and 18% greater during the calibration and verification periods respectively. At the Lemon catchment the flow-period Error Index was lower for both the calibration and verification period. The Nash-Sutcliffe Efficiency was low for the Ernies catchment, probably due to slight over prediction during the onset of streamflow for most of the years (Table 1). The Correlation Coefficients for both the catchments were satisfactory.

6.3.2 Flow duration

The observed and predicted flow-duration curves for the Ernies catchment matched very well (Fig. 13a). The Ernies catchment was flowing only 15% of the time within a year. During the pre-treatment period, Lemon was flowing for a considerably longer period compared to its control. The model over predicted daily streamflow between the range of 0.01 to 5 mm and the flow duration was lower (Fig. 13b). There was a significant change in flow duration of the Lemon catchment following clearing. During 1977–1987 flow-duration increased to 40% which was under predicted by the model. The stream became perennial during 1988–1998 when the permanent groundwater was at the surface. The model predicted the flow duration and the high flows very well but the baseflow was over predicted (Fig. 13b).

6.3.3 Monthly streamflow

The predicted monthly streamflow at the Ernies catchment matched reasonably well with the observed data and the correlations were better than the original monthly model (Bari and Smettem, 2004). The daily model successfully predicted

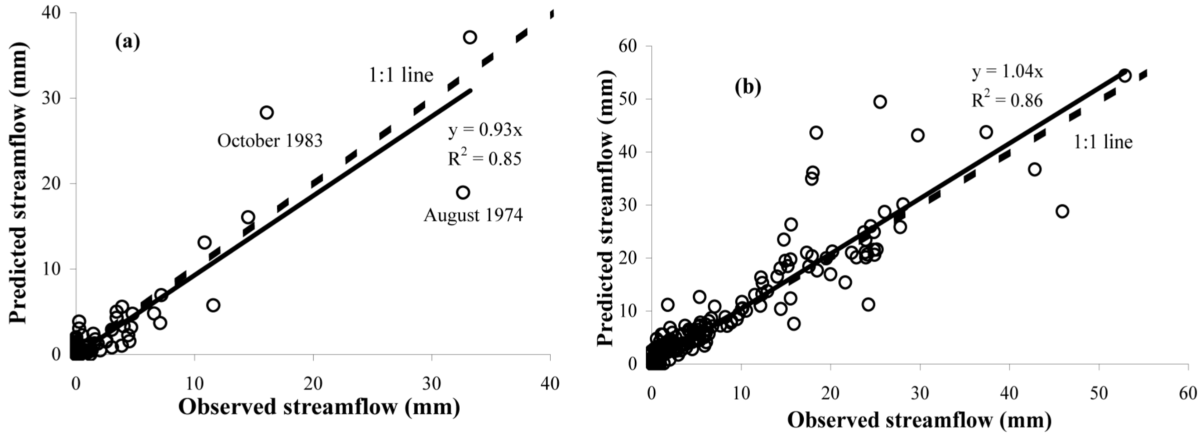

Fig. 14. Monthly flow relationships – **(a)** Ernies and **(b)** Lemon catchments.

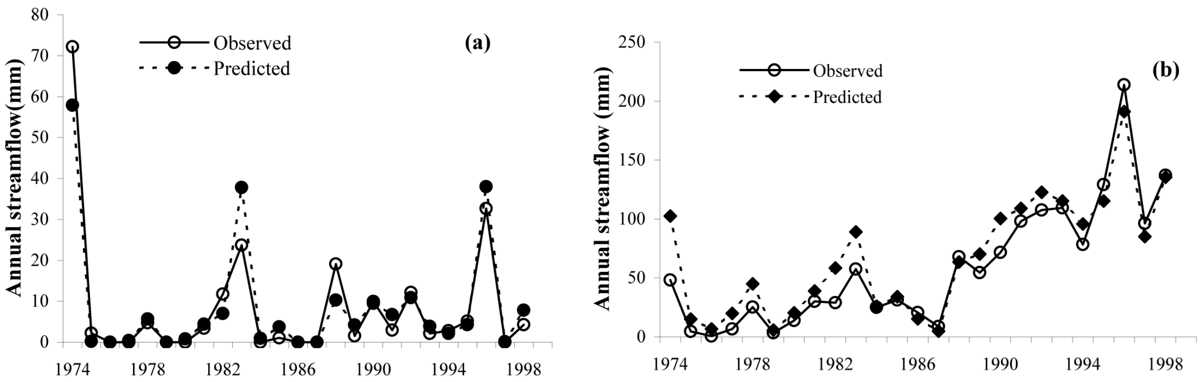

Fig. 15. Annual flow relationships, **(a)** Ernies, **(b)** Lemon catchments.

the January 1982 event and gave improved predictions for other months where the monthly model performed poorly (Fig. 14a). The predicted streamflow for August 1974 was significantly lower than observed while for October 1983, the predicted and observed streamflow was 28 mm and 16 mm respectively. During 1974–1998 there was 7% bias between the observed and predicted monthly streamflow (Fig. 14a). At the Lemon catchment, the daily model over predicted the January 1982 high rainfall event and over predicted for August 1974 when the catchment received the greatest rainfall. When the groundwater system was already at the surface, the model occasionally over predicted the winter high flows (Fig. 14b). Overall, the relationship between the observed and predicted monthly flows was improved by using the daily model rather than the original monthly model. There was 4% bias between the observed and predicted streamflow. Similar monthly relationships were also obtained when the LASCAM model was applied at Wights and Salmon catchments (Sivapalan et al., 1996).

6.3.4 Annual streamflow

At the Ernies catchment, a good agreement between the observed and predicted annual streamflow was observed. In 1974, when the catchment produced the highest streamflow, the model slightly under predicted the streamflow (Fig. 15a). The model generally predicted the low flow years quite well. The observed and predicted flow volumes over the study period were 212 mm and 217 mm respectively. The Ernies catchment lost 12 mm from the groundwater system as downstream discharge and the Subsurface Store decreased by 382 mm, which is comparable to the result of the monthly model (Bari and Smettem, 2004). The storage reduction can be attributed to the reduction in groundwater level, which was observed beneath other forested catchments in the southwest of Western Australia (Schofield and Ruprecht, 1989). The soil evaporation and interception losses were 378 mm and 2932 mm respectively, slightly higher than the prediction of the monthly model (Table 2). The daily model predicted a surface runoff of 75 mm while the monthly model predicted nil, indicating a better representation of daily processes.

Table 2. Water balance components (mm) of Lemon and Ernies catchments (1974–1998).

Component	Sub-component	Lemon	Ernies
Rainfall	Rainfall	17 798	17 798
Evapotranspiration	Interception	1830	2932
	Soil evaporation	1028	378
	Transpiration	11 786	14 571
Storage change	Dry	37	17
	Wet	2	4
	Subsurface	−1343	−380
	Groundwater	2695	50
	Stream zone	48	0
	Loss through base	29	10
Streamflow	Surface runoff	593	75
	Interflow	953	142
	Baseflow	140	0
Mass balance		0	−1

The relationship between the observed and predicted annual streamflow improved significantly compared to that of the annual model (Bari et al., 2005).

During the period of 1974–1984, the model generally slightly over predicted the annual flow observed at Lemon catchment. When the groundwater system came to surface in 1987, the model over predicted the annual streamflow for some of the years (Fig. 15b). Overall, the observed and predicted streamflow was 1469 mm and 1686 mm respectively representing a 15% over prediction. The interception, soil evaporation and transpiration components were 1830 mm, 1028 mm and 11 786 mm respectively (Table 2). The Subsurface Store lost 1343 mm while the Groundwater Store gained 2695 mm due to rise in groundwater level. Similar storage change was also predicted by the annual and monthly models (Bari et al., 2005; Bari and Smettem, 2004). The predicted surface runoff and interflow components were slightly higher while the baseflow was approximately half compared to the prediction of the monthly model. In a related study of daily salinity modelling, we found that the baseflow component is very important of obtaining a good process description.

6.3.5 Sensitivity analysis

Sensitivity analysis of three key model parameters was performed by changing calibrated value one parameter by ±10% at a time. The effects on daily water balance for 1988 at both the catchments were evaluated (Table 3). It demonstrates that the relationship between the moisture content and lateral hydraulic conductivity (ia) is the most sensitive parameter. Reducing 10% increased streamflow at the Lemon and Ernies catchments by 16% and 46% respectively. The daily peakflow increased while the interflow component decreased significantly. Increasing the top soil thickness (d) by 10% resulted in a decrease in streamflow by 9% and 26.6% at the Lemon and Ernies catchment respectively (Table 3). The peakflow component was predicted to decrease while interflow was predicted to increase. The parameter responsible for the spatial distribution of water holding capacity of the top soil (b) has relatively small impact on daily water balance. Further work is necessary across a range of experimental catchments to gain greater insights into the parameter sensitivity.

6.4 General discussion

During 1974–1993, the rainfall at Lemon catchment was about 5% higher than that of Ernies catchment. Since 1984 the rainfall at Lemon catchment was 1% lower than that of Ernies. There was no explanation for this shifting trend in rainfall. Therefore average rainfall obtained from the two catchments was taken as input to the model. During 1974–1983, annual streamflow at Lemon catchment was slightly over predicted (Fig. 15b).

Daily pan evaporation was recorded from Ernies climate station during 1974–1987. There were many gaps in the data and some of the daily data are questionable. The annual pan evaporation data (Luke et al., 1988) was transformed to daily pan evaporation using a simple harmonic function. There was no direct measurement of Leaf Area Index of pasture. A maximum value of 2.1 was used by other models, which were successfully calibrated and tested on Lemon and other similar catchments (Bari and Croton, 2000; Croton and Bari, 2001; Bari and Croton, 2002.). The seasonal variation of LAI was based on the growth pattern of the pasture. There were also no data available for rooting depth and distribution. From field observations and experiences gathered by local farmers, we assumed a maximum pasture rooting depth of 3.0 m. Similar values were also used by other models (Bari and Croton, 2000; Croton and Bari, 2001; Bari and Croton, 2002).

The daily model along with three others were applied to one catchment of the State of Victoria, identified as a National Action Plan for Salinity priority catchment (Beverly et al., 2005). Results show that this model can predict catchment yield using readily available data sets and requires minimum parameterisation and calibration. It can be used as a rapid assessment predictive tool for catchment management and provides subcatchment scale spatial and temporal resolution. It could be used as valuable tool to engage the community and provide a simple representation of relative land use impacts.

Clearing of native forest for pasture development is the major cause of stream salinity and land degradation in Western Australia (Wood, 1924). Other land use changes include forest thinning, operational logging, forest fire and reforestation (Bari and Ruprecht, 2003). A basin scale operational model, which uses the present model as "building block", has now been developed and applied to predict the effects of

Table 3. Sensitivity of selected model parameters for 1988.

Parameter	Relative change(%)	Lemon		Ernies	
		Streamflow (mm)	Error (%)	Streamflow (mm)	Error (%)
ia	+10	75.4	−12.4	14.4	−32.4
	−10	99.8	16.0	31.1	46.0
d	+10	78.3	−9.0	15.6	−26.6
	−10	96.2	11.8	29.1	36.5
b	+10	88.1	2.4	21.6	1.3
	−10	83.9	−2.5	20.9	−2.1

logging, forest fire, clearing and reforestation on streamflow and salinity (Bari and Smettem, 2003; Bari and Berti, 2005). The spatial and temporal variations of land use were incorporated into the input file of the model by changing the proportional area of a subcatchment where land use change took place and changing only the Leaf Area Index and rooting depths. The model has been successfully used for predicting the effects of different rainfall scenarios and climate change on catchment yield (Bari et al., 2005; Bari and Senatherajah, 2005).

7 Summary and conclusion

A conceptual daily model has been developed to represent changes in streamflow generation processes following land use changes and was successfully applied to two experimental catchments in the south-west of Western Australia. The model consists of five inter-connecting stores: (i) Dry, Wet and Subsurface Stores for vertical and lateral water flow, (ii) transient Stream zone Store, and (iii) Groundwater Store. The Dry, Wet and Stream zone Stores represent the dynamically varying stream zone saturated area and are responsible for surface runoff, interflow and percolation. The unsaturated Subsurface Store describes transpiration and quantifies recharge to Groundwater Store. The Groundwater Store quantifies the baseflow and development of the groundwater induced stream zone saturated areas.

The model was calibrated using daily streamflow data. The first 5 years of data (1974–1978) for Ernies catchment and 14 years (1974–1987) of data from Lemon catchment were used for calibration. Catchment average surface slope, soil depth and distribution, porosity, hydraulic conductivity are the most important parameters. Most of the parameters were estimated a priori. Trends in the observed groundwater level were used as a guide to incorporate changes in storage content following forest clearing.

The groundwater level beneath native forest at Ernies catchment remained stable and was successfully reproduced by the model. Streamflow at Ernies catchment is intermittent – flowing generally May to November. The model provided good predictions of the daily streamflow in terms of flow duration, peaks and recessions. During the study period (1974–1998), annual streamflow ranged from nil to 72 mm, averaging 8.5 mm. This represented only 1.2% of annual rainfall. Overall the predicted total streamflow was 2% higher than observed.

At the Lemon catchment the groundwater level rose systematically following clearing and reached the stream bed in 1987. It appears that in the 1990s the groundwater system has reached a new stability. The predicted conceptual groundwater level was in close agreement with the observed data, both beneath the native forest and cleared areas. Following clearing there was a significant increase in streamflow, flow duration, peakflow and recession at Lemon catchment. When the groundwater reached the stream bed in 1987, the annual streamflow increased more than 10 fold. Overall the predicted annual streamflow volume was 15% higher than observed (R^2=0.84).

Appendix A Symbols and variable names

A_i	Total "impervious" area of a catchment (mm^2)
ΔA_i	Changes in 'impervious' area of a catchment (mm^2)
A_p	Pervious area of the catchment (mm^2)
A_s	Part of the catchment area reaching saturation (mm^2)
A_t	Total catchment area (mm^2)
A_w	Part of catchment area where water content exceeded field capacity (mm^2)
a	Parameter for the soil depth distribution of Subsurface Store (-)
b	Parameter for the soil depth distribution of top soil (-)
c	Parameter for the soil depth distribution of top soil(-)
C_i	Interception store coefficient (-)
C_{loss}	Parameter for Groundwater Store loss (-)
cs	Parameter related to soil evaporation (-)
C_s	Plant canopy storage (mm)
C_{smx}	Maximum interception storage capacity (mm)
d	Average depth of top soil (mm)
d'	Depth of an elementary area of top soil (mm)
d_l	Total depth of the soil profile (mm)
d_g	Average depth to groundwater level (mm)
d_s	Stream depth (mm)
d'_g	Depth to groundwater level of any elementary area (mm)
Δd_g	Changes in groundwater level (mm)
E_{sd}	Soil evaporation from Dry Store (mm)
E_{ss}	Soil evaporation from Stream zone Store (mm)
E_{sw}	Soil evaporation from Wet Store (mm)
E_{td}	Actual transpiration from Dry Store (mm)
E_{tg}	Actual transpiration from Groundwater Store (mm)
E_{tl}	Actual transpiration from Subsurface Store (mm)
E_{ts}	Actual transpiration from Stream zone Store (mm)
E_{tw}	Actual transpiration from Wet Store (mm)
I	Percolation (mm)
I_a	Actual interception (mm)
ia	Parameter related to lateral conductivity of top soil (-)
K_{ll}	Lateral hydraulic conductivity of Subsurface Store (mm day^{-1})
K_{lv}	Vertical hydraulic conductivity of the Subsurface Store (mm day^{-1})
K_{ul}	Lateral hydraulic conductivity of West Store (mm day^{-1})
K_{uv}	Vertical hydraulic conductivity of Wet Store (mm day^{-1})
L	Catchment wide average stream length (mm)
LAI	Leaf Area Index (-)
LAI_{mx}	Maximum Leaf Area Index (-)
pa	Parameter related to vertical soil conductivity (-)
pb	Parameter related to vertical soil conductivity (-)
PET	Daily pan evaporation (mm)
Q_i	Interflow (mm)
Q_r	Total surface runoff (mm), ($Q_{r1}+Q_{r2}$)
Q_{r1}	Surface runoff from pervious area (mm)
Q_{r2}	Surface runoff from "impervious area"(mm)
Q_b	Baseflow to stream (mm)
Q_{bl}	Baseflow to Stream zone Store (mm)
Q_{loss}	Groundwater loss below gauge (mm)
Q_t	Total streamflow (mm)
R	Actual rainfall (mm)
RE	Effective rainfall (mm)
RET	Residual potential evapotranspiration (mm)
Rf	"Excess water" released from Dry Store to Wet Store (mm)
R_l	Recharge to Groundwater Store (mm)
R_{l1}	Recharge to Groundwater Store by matrix flow (mm)
R_{l2}	Recharge to Groundwater Store by preferential flow (mm)
RT_t	Total root volume (-)
RT_u	Root volume in the top soil (-)
tu	Parameter related to transpiration (-)
W_d	Water content of the Dry Store (mm)
W_{dmx}	Maximum capacity of the Dry Store (mm)
W_g	Water content of the Groundwater Store (mm)
ΔW_{gl}	Changes in water between Subsurface and Groundwater Stores (mm)
W_l	Water content of the Subsurface Store (mm)
W_{ldmx}	Water content at field capacity of the Subsurface Store (mm)
W_{lmx}	Maximum capacity of the Subsurface Store (mm)
ΔW_l	Changes in water content of the Subsurface Store (mm)
W_{sg}	Water content of the Stream zone Store (mm)
ΔW_{sg}	Changes in water content of the Stream zone Store (mm)
W_w	Water content of the Wet Store (mm)
W_{wi}	Threshold value for interflow generation (mm)
W_{wmx}	Maximum capacity of the Wet Store (mm)
w_d	Elementary area water retention capacity of Dry Store (mm)
w_{dm}	Dry Store maximum water retention capacity of an elementary area (mm)
w_l	Water holding capacity of the subsurface elementary area (mm)
w_{lm}	Maximum water holding capacity of any subsurface elementary area (mm)
w_w	Elementary area water retention capacity of the Wet Store (mm)
w_{wm}	Wet Store maximum water retention capacity in any elementary area (mm)
x	Ratio of the "pervious" area of top soil which exceeded field capacity (-)
x'	Ratio of the subsurface unsaturated area which exceeded field capacity (-)
y	Ratio of the "pervious" area of a catchment which reached saturation (-)
α_t	Parameter related to transpiration (-)
β	Catchment average groundwater slope (mm/mm)
θ_s	Water content at saturation of an elementary area (mm^3 mm^{-3})
θ_f	Field capacity of top soil elementary area (mm^3 mm^{-3})
θ_{lf}	Field capacity of subsurface elementary area (mm^3 mm^{-3})
ϕ_l	Soil porosity of the subsurface elementary area (mm^3 mm^{-3})

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