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Numerical modelling of water transfer among precipitation, surface water, soil moisture and groundwater

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Abstract In the course of the hydrological cycle, when precipitation reaches the ground surface, water may become surface runoff or infiltrate into soil and then possibly further percolate into the groundwater aquifer. A part of the water is returned to the atmosphere through evaporation and transpiration. Soil moisture dynamics driven by climate fluctuations play a key role in the simulation of water transfers between the ground surface, unsaturated zone and aquifer. In this study, a representation of one canopy layer and four soil layers is used for a coupled soil-vegetation modelling scheme. A non-zero hydraulic diffusivity between the deepest soil layer modelled and groundwater table is used to couple the numerical equations of soil moisture and groundwater dynamics. Simulation of runoff generation is based on the mechanism of both infiltration excess overland flow and saturation overland flow nested in a numerical model of soil moisture dynamics. Thus, a comprehensive hydrological model integrating canopy, soil zone and aquifer has been developed. The model was applied to simulate water transfers between precipitation, surface water, soil moisture and groundwater for assessing water resources in the plain region of the Huaihe River basin in east China. The newly developed model is capable of calculating hydrological components of surface runoff, evapotranspiration from soil and aquifer, and groundwater recharge from precipitation and discharge into rivers. Regional parameterization was carried out by using two approaches. One is to determine the majority of parameters representing specific physical values on the basis of characterization of soil properties in the unsaturated zone and aquifer, and vegetation. The other is to calibrate the remaining few parameters on the basis of comparison between measured and simulated streamflow and groundwater table. The integrated modelling system was successfully used in the Linhuanji catchment of the Huaihe plain region. Study results show that: (a) on the average 14.2% of precipitation becomes surface runoff and baseflow during a 10-year period from 1986 to 1995 and this figure fluctuates between only 3.0% in dry years of 1986, 1988, 1993 and 1994 to 24.0% in the wet year of 1991; (b) groundwater directly deriving from precipitation recharge is about 15.0% of the precipitation amount, and (c) about half of the groundwater recharge flows into rivers and losses through evaporation.

Key words coupled soil–vegetation modelling; groundwater dynamics; Huaihe River, China; numerical modelling; soil moisture

INTRODUCTION

Atmospheric, surface and subsurface portions of the hydrological system are dynamically linked water reservoirs which are characterized at different time and space scales. Many challenges remain in understanding and evaluating the dynamic interactions among these reservoirs, especially for those between surface and subsurface layers (NRC, 2004). In engineering and Earth science investigations, there has been a long tradition of assigning soil moisture below the root zone to be groundwater, which is used as a "boundary condition" in the simulation of soil moisture dynamics. In river hydraulics and hydrodynamics of open channels, the porous subsurface is seldom considered to be an active participant of in-channel processes and dynamics. In atmospheric science, soil moisture and groundwater are represented as "buckets" of limited size and dynamics uncoupled to rivers. To other scientists and resource managers, groundwater is represented as an infinitely large and slow process, which is unlikely to play a significant role in the hydrological process at the time scales of regular human activities (Duffy, 2004).

Regional water resources include surface water and groundwater, and quantitative estimation of water resources should be based on simulation of rainfall–runoff processes, precipitation recharge and water loss from evapotranspiration. Traditionally, conceptual hydrological models established simple relationships of the rainfall–runoff and the evaporation-soil moisture loss (Zhao, 1980), which are widely used for water resources estimation and water transfer among surface water, soil moisture and groundwater in China (Shen, 1992; Xu & Guo, 1994; Guo *et al.*, 1997). The atmosphere, land cover and human activity are the driving force and important variable respectively for the changes in the hydrological cycle. The records of river runoff and groundwater table are usually the integrated results of all changes in atmosphere, catchment surface and the artificial direct uses of river water and groundwater. For quantitative evaluation of these changes individually, a distributed physically-based hydrological model is needed since it can represent the spatial distribution of related basin properties and can examine the impacts of local changes on the basin hydrological cycle.

This study focuses on the development of a comprehensive hydrological modelling system in the plain region of the Huaihe River catchment where they actually face the dilemma of having to undergo rapid economic development and greater water demands for agriculture and industry. Because water resources are stressed, hastily and rather inefficiently exploited in this region, the wisdom of past generations in managing water got lost. The modelling system can be helpful for sustainable water resources management by recognizing the hydro-system complexity and interconnectivity of its elements. The model was applied in the Linhuanji catchment of the Huaihe River region and is primarily used for estimation of surface runoff, precipitation recharge, and groundwater losses for evapotranspiration and stream baseflow. In the next section, the model structure of a multi-layer soil moisture model coupling the two-dimensional (2-D) groundwater was introduced. In the following sections, the integration scheme based on numerical method and interface water fluxes was developed for the estimation of surface runoff, precipitation recharge into groundwater and evapotranspiration, and the integrated model was applied in the Linhuanji catchment of the Huaihe River plain region (Fig. 1). The model parameters were determined based on hydrological properties of soil and aquifer, and calibration of observed streamflow discharges and groundwater levels. The regional water budget was calculated for estimation of available water resources. The final section gives the conclusion.

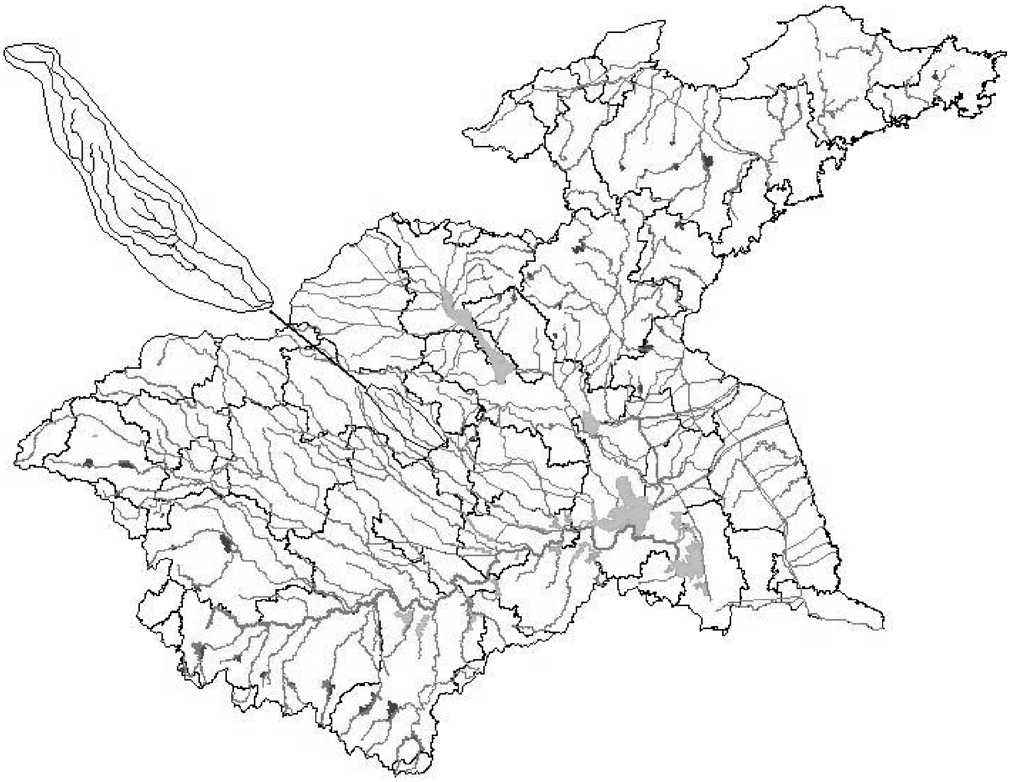


Fig. 1 Plain region of the Huaihe River watershed and location of the study site.

MODEL STRUCTURE

Soil moisture

Soil moisture variation in the model is described by Richards' equation. Integrating Richards' equation through four soil layers under the assumption of vertically homogeneous soil hydraulic properties within each layer yields the following equations (Chen & Duhia, 2001; Chen & Hu, 2004):

$$d_1 \frac{\partial \theta_1}{\partial t} = -D\left(\frac{\partial \theta}{\partial z}\right)_1 - K_1 + P_d - R_s - E_{dir} - E_{T1} \tag{1}$$

$$d_2 \frac{\partial \theta_2}{\partial t} = D\left(\frac{\partial \theta}{\partial z}\right)_1 - D\left(\frac{\partial \theta}{\partial z}\right)_2 + K_1 - K_2 - E_{T2} \tag{2}$$

$$d_3 \frac{\partial \theta_3}{\partial t} = D\left(\frac{\partial \theta}{\partial z}\right)_2 - D\left(\frac{\partial \theta}{\partial z}\right)_3 + K_2 - K_3 - E_{T3} \tag{3}$$

and

$$d_4 \frac{\partial \theta_4}{\partial t} = D\left(\frac{\partial \theta}{\partial z}\right)_3 - D\left(\frac{\partial \theta}{\partial z}\right)_4 + K_3 - K_4 \tag{4}$$

where subscript i = 1, 2, 3, and 4 is soil layer index, d_i is thickness of the i-th soil layer, P_d is precipitation falling on the ground, R_s is surface runoff, K_i is vertical unsaturated soil hydraulic conductivity and D the soil water diffusivity. Both K and D are functions of soil moisture content θ and are computed from $K(\theta) = K_s(\theta/\theta_s)^{2b+3}$ and $D(\theta) = K(\theta)(\delta\psi/\delta\theta)$, where ψ is soil water tension function and $\psi(\theta) = \psi_s/(\theta/\theta_s)^b$ in which b is a curve-fitting parameter. Equation (4) includes upward soil moisture transfer between the deepest model soil layer and the groundwater table.

Evapotranspiration

In soil moisture model (SMM), the total evaporation, ET, is the sum of: (1) direct evaporation from the top shallow soil layer, E_{dir}; (2) transpiration via canopy and roots, E_t; and (3) evaporation of precipitation intercepted by the canopy, E_c.

A simple linear method is used to calculate E_{dir} (Mahfouf & Noilhan, 1991):

$$E_{dir} = (1 - \sigma_f)\beta E_p \tag{5}$$

where $\beta = \dfrac{\theta_1 - \theta_w}{\theta_{ref} - \theta_w}$, in which θ_{ref} and θ_w are field capacity and wilting point, respectively. E_p is the potential evaporation calculated using a Penman-based energy balance approach that includes a stability-dependent aerodynamic resistance (Mahrt & Ek, 1984), and σ_f is the green vegetation fraction (cover). E_t is calculated by:

$$E_t = \sigma_f E_p B_c \left[1 - \left(\frac{W_c}{S} \right)^n \right] \tag{6}$$

where B_c is a function of canopy resistance, and W_c is intercepted canopy water content, which is calculated according to the budget for intercepted canopy water, and S is the maximum canopy capacity and n = 0.5. Finally, the third component of the total evaporation, E_c, can be estimated by:

$$E_c = \sigma_f E_p \left(\frac{W_c}{S} \right)^n \tag{7}$$

The budget for intercepted canopy water is:

$$\frac{\partial W_c}{\partial t} = \sigma_f P - P_d - E_c \tag{8}$$

where P is total precipitation. If W_c exceeds S, the excess precipitation or drip, P_d, reaches the ground.

SIMULATION OF RUNOFF

Surface runoff

In the semihumid region of China, infiltration excess overland flow and saturated overland flow can be generated from precipitation. The former surface runoff, R_s, is defined as the excess of precipitation which does not infiltrate into the soil ($R_s = P_d - I_{\max}$). The maximum infiltration, $I_{\max}$, is formulated as:

$$I_{\max} = \min\{K_1, I_f\} \tag{9}$$

where K_l is the upper layer soil hydraulic conductivity and I_f is the infiltration capacity related to precipitation intensity, soil moisture deficit and rainfall duration (Chen & Dunhia, 2001).

In the wet season, the upper layer soil may be saturated during a rainfall event, resulting in overland flow ($R_s = \max\{P_d - D_{x1}, 0\}$; D_{x1} is the upper layer soil moisture deficit).

Surface runoff is routed by a time lag approach representing the watershed regulation and the channel system to the stream outlet. The calculation equation is:

$$Q_s(t) = CS * Q_s(t - \Delta t) + (1 - CS) * \overline{R}_s(t - Lag) \tag{10}$$

where $Q_s(t)$ and $Q_s(t - \Delta t)$ and are the outlet discharges by surface runoff at time t and $t - \Delta t$, respectively; $\overline{R}_s$ is average value of surface runoff R_s between time t and $t - \Delta t$; CS is coefficient of surface runoff concentration; Lag is time lag.

Groundwater

A portion of the precipitation recharge to groundwater flows into the stream as baseflow. The flow rate Q_g between stream channel and aquifer is calculated from the difference in hydraulic heads in the stream and the adjacent aquifer using the following equation (McDonald & Harbaugh, 1988):

$$Q_g = C_{riv}(H_{riv} - h) \tag{11}$$

where Q_g is the flow between the stream and the aquifer, H_{riv} is the head in the stream channel, h is the head at the node in the cell underlying the stream reach, C_{riv} is the hydraulic conductance of the stream–aquifer interconnection. Baseflow, the recharge and groundwater evapotranspiration depend on the level of groundwater table, which is described by the governing equation in a 2-D form:

$$\frac{\partial}{\partial x}(Kh\frac{\partial h}{\partial x}) + \frac{\partial}{\partial y}(Kh\frac{\partial h}{\partial y}) = S_y\frac{\partial h}{\partial t} - W_s \tag{12}$$

where S_y is specific yield; W is a volumetric flux per unit volume representing sources and/or sinks of water, including equation (11) and (13), with $W > 0$ and $W < 0$ for flow in and out of the groundwater system, respectively. The finite-difference groundwater model MODFLOW (McDonald & Harbaugh, 1988; Harbaugh *et al.*, 2000) was used for solving equation (12).

Water exchanges between unsaturated and saturated zones

Precipitation recharge into groundwater or groundwater loss from evapotranspiration is water exchange on the interface between saturated and unsaturated zones. The exchange rate W_e can be estimated by the following equation:

$$W_e = K(\Psi)\left(\frac{\partial \Psi}{\partial z} - 1\right)_4 = D\left(\frac{\partial \theta}{\partial z}\right)_4 - K_4 \tag{13}$$

where $D\left(\frac{\partial \theta}{\partial z}\right)_4 = D\frac{\theta_s - \theta_4}{Z_g}$, and Z_g is the distance between groundwater table and the mid-point of the affected layer. W_e is the recharge (drainage from the vadose zone) P_{rg} or groundwater loss to the soil through evapotranspiration E_g.

Integration of numerical models

The equations for simulation of soil moisture and groundwater dynamics due to precipitation infiltration and evapotranspiration are coupled for calculation of water transfer among precipitation, soil moisture, surface water and groundwater. The coupling is based on numerical approaches by discretizing the whole catchment into grids, each of which is considered to be hydrologically and hydrogeologically uniform. Vertically, coupling of soil moisture dynamics and groundwater flow is based on the water exchange at the interface between the unsaturated and saturated zones using equation (13). Surface water and groundwater interaction is based on equation (11). The whole modelling system is shown in Fig. 2. It illustrates that after precipitation is intercepted by vegetation, a portion of it, P_d, reaches the ground surface. Water may become surface runoff, R_s, or infiltrate into soil and then further percolate into the groundwater aquifer, P_{rg}. A part of the water is returned to the atmosphere through evaporation and transpiration, *ET*. Groundwater may supply upper layer soil moisture through evapotranspiration, E_g. The inputs and outputs of the unsaturated and saturated zones lead to variations of soil moisture and groundwater tables, which can be simulated by the soil moisture model in equations (1)–(4) and groundwater model in equation (12). Additionally, the model includes artificial influences to the water exchanges, e.g. groundwater pumping for irrigation and artificial ponds for storing water.

The model requires a number of soil parameters including saturated hydraulic conductivity, water contents at wilting point, field capacity and saturation. They are physically-based parameters and can be determined primarily regarding soil properties in this study. The other parameters without physical significance, such as coefficient of surface runoff concentration, *CS*, time lag, storage capacity of small ponds and hydraulic conductivity, C_{riv}, need to be calibrated on the basis of observed streamflow discharge and groundwater table.

Model execution strategies are: (1) the multi-layer soil moisture model is used for simulating infiltration, recharge, actual ET, surface runoff, and soil moisture in the root zone in each grid by using specification of rainfall, reference ET, and crop characteristics; (2) surface runoff concentration is further routed by the time lag approach and the channel system to the stream outlet for calculation of watershed outlet discharge. The interchanges of stream channel flow and groundwater flow are calculated by River Package of MODFLOW in equation (11); (3) estimations of precipitation recharge, groundwater evapotranspiration loss and baseflow from

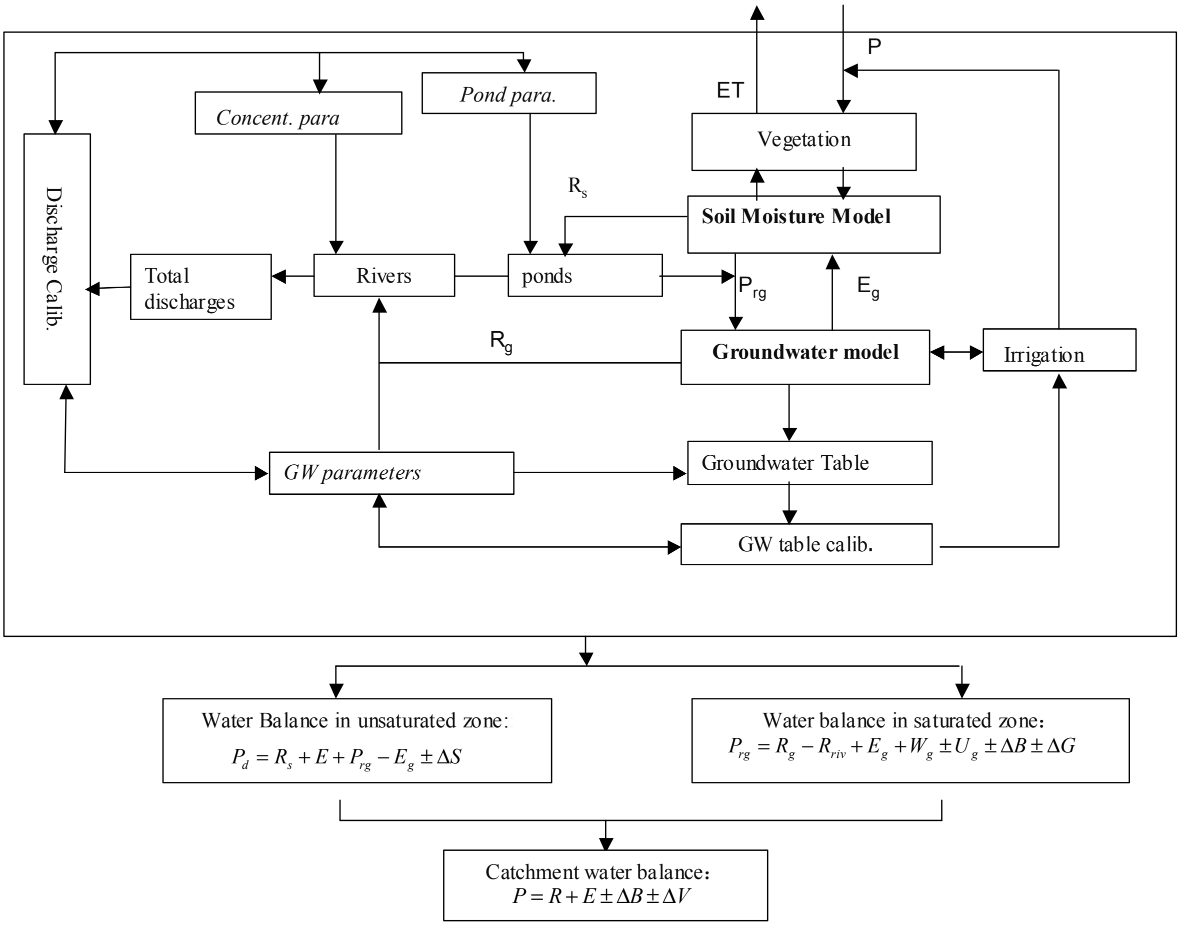


Fig. 2 Scheme of model execution.

SSM, equation (13) and (11), respectively, were used to alter the MODFLOW calculation from Recharge Package, Evapotranspiration Package and River Package. USGS FORTRAN program of MODFLOW-2000 was applied for groundwater flow; (4) the model will be calibrated and validated against measured groundwater elevations, stream flow discharges, and water balances in the unsaturated zone and saturated zone. A trial and error method was used for model calibration. SSM and MODFLOW are iteratively executed until good matches of observed and simulated stream flow discharges, and observed and simulated groundwater tables are obtained.

APPLICATIONS

Model calibration and validation

The model was applied in the Linhuanji catchment within a semihumid or semiarid region of the Huaihe River watershed (Fig 1). The catchment area is 2560 km^2. Average annual precipitation during 1986–1995 was 713 mm, and approximately 60~70% of the precipitation fell in summer season from June to September. The annual potential evapotranspiration is 960 mm.

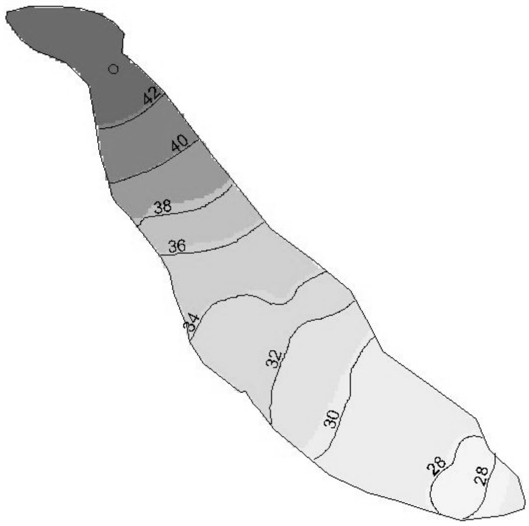


Fig. 3 Ground surface elevation in the study site.

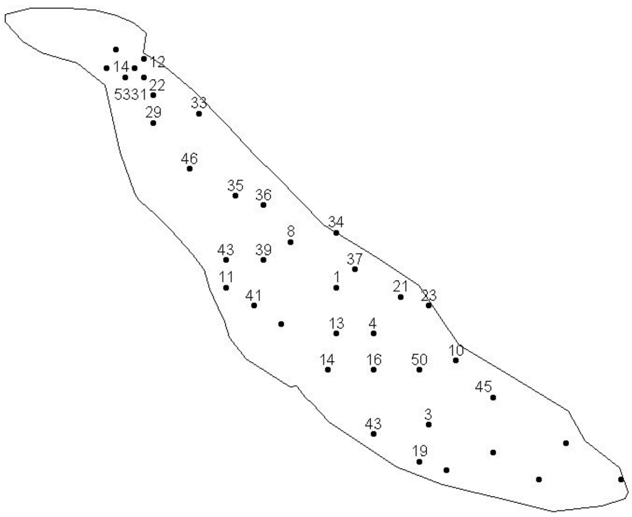


Fig 4 Location of groundwater observation wells.

Spatial variations of groundwater table, similar to the ground surface elevation (Fig. 3), are from 45 m in the north to 28 m in the south. Groundwater table is as deep as 7–8 m in the north and its depth decreases by 2 m in the south. Annual groundwater fluctuations are approximately 1~2 m.

For estimation of water transfer in the region, we collected 10 years of data from 1986 to 1995, including daily precipitation of 25 observation stations, pan evaporation, groundwater table at a five day interval from 30 observation stations (Fig. 4), and daily streamflow discharge at the catchment outlet. Spatial distribution of soil properties and vegetation are also available. The study region is located in the alluvial plain of Yellow River, and overlaid by loose deposits of sand and silty sand (Fig. 5). In the upper cultivated soil of the unsaturated zone, the soil is sandy loam in the north and sandy clay loam in the south. Wheat, maize and sorghum are the main crops in the region.

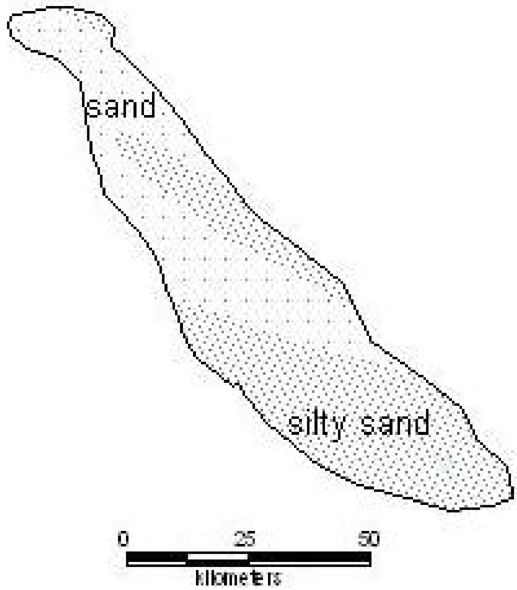
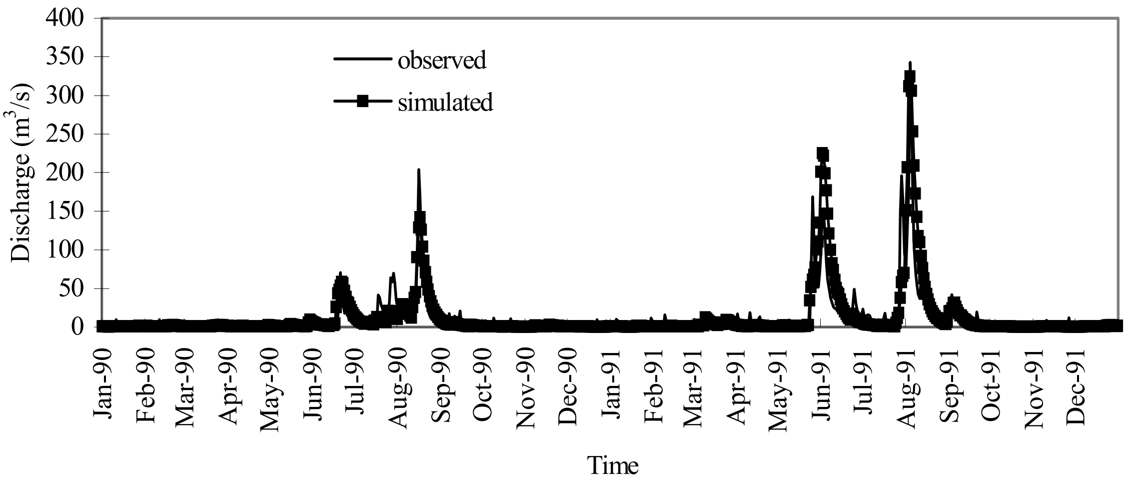


Fig 5 Geological material of aquifer in the study site



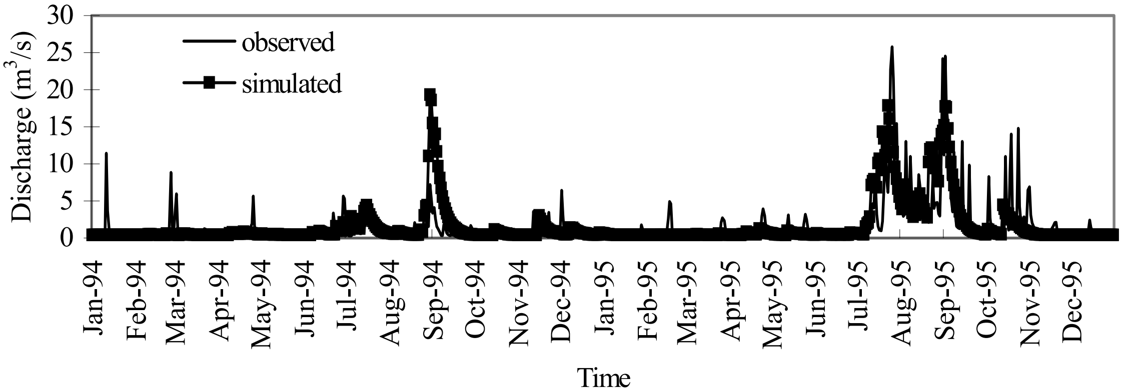


Fig 6 Observed and simulated discharges.

The study region was discretized into 2356 grid units, each 1047 m long and 1048 m wide. The simulation time step is one day. The physical parameters of soil moisture dynamics in the unsaturated zone are specified by the soil analysis of Cosby *et al.* (1984), and hydraulic conductivities are 0.41 and 0.29 m day^{-1} in the sandy loam and sandy clay loam, respectively. In the saturated zone, hydraulic conductivity and specific yield in the sand region are 4.4 m day^{-1} and 0.055, respectively, and 2.8 m day^{-1} and 0.045 in the silty sand region, respectively. Other parameters were calibrated from observed stream discharge and groundwater table by the trial and error method. The calibrated coefficient of surface runoff concentration *CS* and time lag is 0.35 and 1 day, respectively; storage capacity of small ponds is 2.5 mm. Simulated and

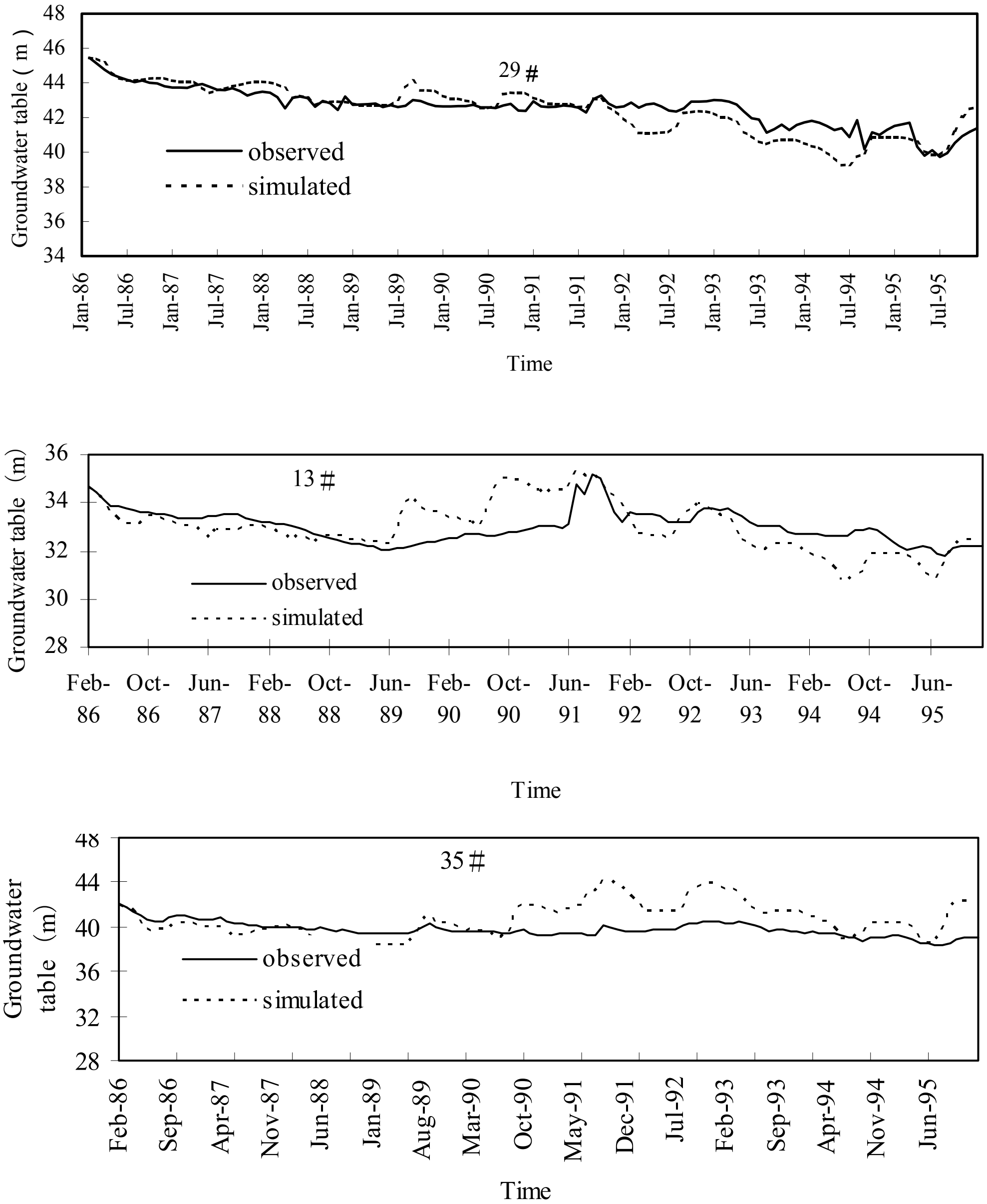


Fig 7 Simulated and observed groundwater table at observation wells 29, 13 and 35.

observed discharge, 1990~1991 and 1994~1995 as examples, are shown in Fig. 6. Simulated stream flow discharge usually represents the observations quite well. The Nash-Sutcliff index for 1986–1995 is 0.79 and RMSE is 0.4 $m^3\ s^{-1}$. Some larger errors arose from artificial influences, such as numerous ponds and dams, pumping for irrigation, in that observation data for accurate estimation of their storage capacities and water uses are difficult. Hydraulic conductivity C_{riv} was determined by hydraulic conductivity of riverbed deposits and river cross-section features. It ranges from 200 to 1500 $m^2\ day^{-1}$. The flows into and out of the aquifer were further validated by using observed groundwater tables. Figure 7 demonstrates the simulated and observed groundwater tables for observation wells of number 29, 13 and 35. Figure 8 is average values of the simulated and the observed groundwater tables for 30 observation wells. The simulated groundwater table generally matches the observed groundwater table well. The correlation between them is 0.81 (Fig. 9).

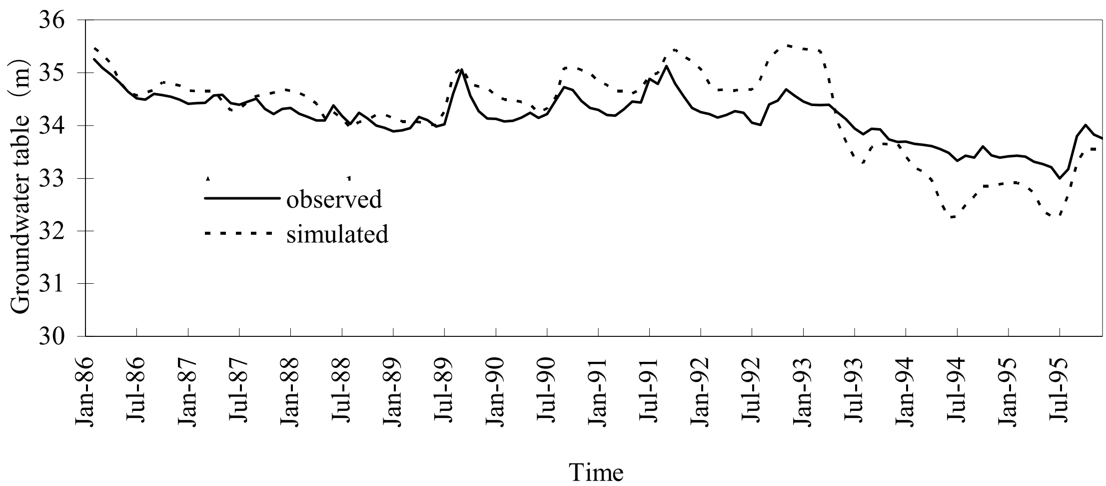


Fig 8 Average values of simulated and observed groundwater table for 30 observation wells.

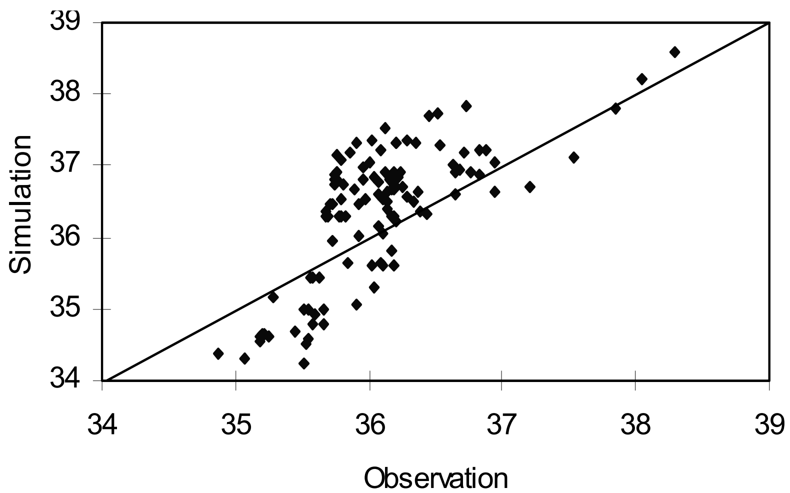


Fig. 9 Correlation between area-average observed and simulated groundwater tables.

Water budget

Table 1 shows the simulation results of precipitation transferring into the recharge, surface runoff, total streamflow and baseflow, and evapotranspiration loss. The calculated runoff is the total of surface runoff R_s and baseflow R_g. The yearly relative errors between observed and calculated runoff range between 1.94 and 22.84. Larger simulation errors occur primarily in the drought years because the artificial influences become more intensive for meeting the water requirements. For the multi-year averages, the calculated runoff is very close to the observation.

The simulated water budget items by the model in Table 1 and 2 clearly offer information for water transfers from precipitation to available surface runoff and groundwater, and water losses for evapotranspiration in the study region. For the 10-year average, the total of yearly mean precipitation and groundwater withdrawal primarily for agricultural utilization (columns (2) + (8) in Table 1) is 769.46 mm, and actual evaporation from soil moisture and vegetation transpiration is 669 mm (column (3)), approximately 87% of the total of precipitation and groundwater withdrawal amount. The yearly mean of runoff is 58 mm (column (9)), approximately 8.2% of precipitation amount (i.e. annual mean runoff coefficient equals 0.082). It varies from 3.0% in the dry years of 1986, 1988, 1993 and 1994 to 24.0% in the wet year of 1991. The yearly mean of precipitation recharge into groundwater is 103.6 mm (column (6)), approximately 15% of precipitation amount (i.e. annual mean recharge coefficient equals 0.15). It varies from 3.0% to 20.0%. More than half of the recharge amount losses for evapotranspiration and flows into stream channel as baseflow (column (4) and column (7)), and the remaining amount is stored in the aquifer.

Table 2 shows multiyear average water balances during 1986–1995 in unsaturated and saturated zones. For unsaturated zone, the inputs include precipitation P and irrigation water from groundwater withdrawal, and upward flow of groundwater through evaportranspiration E_g, and the outputs include soil moisture loss through evaporation and transpiration E, surface runoff R_s and precipitation recharge to groundwater P_{rg}. The relative error of water balances $\frac{Inputs - Output}{storage\,changes}$ is approximately 1%. For the

Table 1 Water budget and relative errors between observed and calculated runoff.

Year (1)	P (mm) (2)	Evapotrans–piration (mm) (3)	E_g (mm) (4)	R_s (mm) (5)	P_{rg} (mm) (6)	Baseflow R_g (mm) (7)	GW withdrawal, W_g (mm) (8)	Cal. runoff (mm) (9)	Obs. runoff (mm) (10)	Relative error (%) (11)
1986	571	648	64.93	21.45	47.38	8.70	63.06	30.16	32.36	7.32
1987	713	653	48.37	18.19	65.25	9.00	31.39	27.19	27.72	1.94
1988	572	723	39.06	8.60	63.63	16.22	74.20	24.82	29.91	20.50
1989	813	647	49.83	165.49	28.47	41.64	26.17	90.84	104.5	15.04
1990	887	667	69.46	73.13	163.78	19.32	39.46	92.45	96.25	4.12
1991	883	638	86.43	168.49	174.74	39.88	60.91	208.4	195.91	–5.98
1992	751	726	46.11	42.67	111.16	12.70	49.91	55.37	48.52	–12.38
1993	616	593	25.29	12.97	29.79	3.38	79.03	16.35	17.12	4.69
1994	610	716	6.93	8.64	71.72	3.18	58.74	11.82	10.36	–12.38
1995	723	684	19.06	15.65	143.01	5.55	56.29	21.20	26.04	22.84
Mean	714	669	45.55	43.22	103.60	14.64	55.46	57.86	58.88	1.75

Table 2 Water balance in unsaturated and saturated zones during 1986–1995.

Unsaturated zone					
Input (mm)	*P*	E_g	Irrigation	Storage changes	Balance error (%)
	713.82	45.55	42.0	17.7	
Output (mm)	Evapotrans-piration E	*Rs*	P_{rg}		
	669.45	43.22	103.60		–1.0
Saturated zones					
Input (mm)	P_{rg}			Boundary incomes (input-output)	Balance error (%)
	103.60			12.06	
Output (mm)	Baseflow R_g	GW withdrawal, W_g	E_g		
	14.64	55.46	45.55		0

saturated zone, the recharge from the bottom of the unsaturated zone and the net inflow from the boundary is the aquifer input. A portion of the recharge flows into the stream as baseflow or losses through evapotranspiration and withdrawal for irrigation. The water budget in the unsaturated and saturated zones indicates that the model simulation is able to keep track of water balance well in the study region.

CONCLUSIONS

A comprehensive modelling system based on soil moisture and groundwater dynamics in numerical solutions has been developed for simulation of the hydrological processes of precipitation recharge, surface water, groundwater, and soil moisture content and groundwater table. The model has been successfully applied in the plain area of Linhuanji catchment. Water balance simulation was based on spatial data of topography, soil and plants, meteorological data and hydrological data of stream discharge and groundwater tables. The model is very useful for water resources assessment and planning. The capability of soil moisture content prediction enables the model to be useful tool for scheduling agricultural irrigation planning.

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