Numerical Modeling of Water Transfer among Precipitation, Surface Water, Soil Moisture and Groundwater

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Abstract : In the processes of hydrological cycle, when precipitation reaches the ground surface, water may become surface runoff or infiltrate into soil and then possibly further percolate into groundwater aquifer. A part of the water is returned to the atmosphere through evaporation and transpiration. Soil moisture dynamics driven climate fluctuations plays a key role in the simulation of water transfer among ground surface, unsaturated zone and aquifer. In this study, a one-layer canopy and a four-layer soil representation is used for a coupled soil-vegetation modeling scheme. A non-zero hydraulic diffusivity between the deepest soil layer modeled 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 to evaluate water resources in the plain region of Huaihe River basin in East China and simulate water transfer among precipitation, surface water, soil moisture and groundwater. The newly developed model is capable of calculating hydrological components of surface runoff, evapotranpiration from soil and aquifer, and groundwater recharge from precipitation and discharge into rivers. Regional parameterization is made by using two approaches. One is to determine most parameters representing specific physical values on the basis of characterization of soil properties in unsaturated zone and aquifer, and vegetations. The other is to calibrate the remaining few parameters on the basis of comparison between measured and simulated streamflow and groundwater tables.

The integrated modeling system was successfully used in the Linhuanji catchment of Huaihe plain region. Study results demonstrate that (1) on the average 14.2% of precipitation becomes surface runoff and baseflow during a ten-year period from 1986 to 1995 and this figure fluctuates between only 3.0% in drought years of 1986, 1988, 1993 and 1994 to 24.0% in wet year of 1991; (2) groundwater directly deriving from precipitation recharge is about 15.0% t of the precipitation amount, and (3) about half of the groundwater recharge flows into rivers and loses through evaporation.

1 Introduction

Atmospheric, surface and subsurface portions of the hydrological system are dynamically linked water reservoirs having distinctly different time and space scales. Many challenges remain in understanding and measuring the dynamic interchange among these reservoirs, especially for interchanges with the subsurface (NRC, 2004).

Lack of a coherent strategy for integrated observations of soil-moisture-groundwater level and streamflow for the nations watersheds and river basins must await a scientific investigation to assess the degree to which the subsurface plays an “active” role in the dynamics of the land-surface-river continuum from hillslope to watershed to river-basin scales. There has been a long tradition in engineering and earth science investigations of assigning soil moisture below the root zone and groundwater the role of a passive “boundary condition”. In the case of river hydraulics and the hydrodynamics of open channels, the porous subsurface is rarely considered to be an active participant of in-channel processes and dynamics. In the atmospheric sciences, soil moisture and groundwater have been represented as “buckets” of limited size and dynamics uncoupled to rivers. To other scientists and resource managers, groundwater has represented an infinitely large and slow process unlikely to participate over human time scales at all (Duffy, 2004).

Estimation of water resources should be based on integrated hydrologic modeling system. 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 (Shen, 1992; Guo et al., 1997; Xu and Guo,1994). For investigating water resources in artificially influenced watersheds, distributed hydrologic models are developed in recent years. Hydrologic processes and its climate and artificial variations are analyzed in detail.

This study is focus on development of a comprehensive hydrologic modeling system in plain region. The model was applied in the Linhuanji catchment of Huaihe plain region and is primarily used for estimation of water resources and water transfer among precipitation, surface water, soil moisture and groundwater.

2 Model structure

2.1 Soil moisture

Soil moisture variation in the model is described by the Richard’s equation. Integrating the Richard’s equation through four soil layers under the assumption of vertically homogeneous soil hydraulic properties with each layer yields

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

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

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

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 the soil layer index, d_i is the thickness of *i*th soil layer, P_d the precipitation falling on the ground, R the surface runoff, K_i the vertical unsaturated soil hydraulic conductivity. Eq (4) includes upward soil moisture transfer between the deepest model soil layer and the groundwater table.

2.2 Evapotranspiration

In SMM, the total evaporation, *ETa,* is the sum of 1) the direct evaporation from the top shallow soil layer, E_{dir}; 2) evaporation of precipitation intercepted by the canopy, *Ec*; and 3) transpiration via canopy and roots, E_t . That is, $ETa = E_{\text{dir}} + E_c + E_t$.

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

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

where $\beta = \dfrac{\theta_1 - \theta_w}{\theta_{ref} - \theta_w}$, in which θ_{ref} and θ_w are the field capacity and wilting point, respectively. EP is the potential evaporation calculated by a Penman – based energy balance approach that includes a stability-dependent aerodynamic resistance (Mahrt and 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 the 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. In addition, the total evapotranspiration includes evaporation of precipitation intercepted by the canopy, E_c

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

The budget for intercepted canopy water is

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

Where *P* is the input total precipitation. If *Wc* exceeds *S,* the excess precipitation or drip, *D,* reaches the ground.

2.3 Runoff Calculation

2.3.1 Surface runoff

In the semi-humid region of China, infiltration excess and saturated runoff could be formed for a precipitation. The former surface runoff, *R,* is defined as the excess of precipitation not infiltrated into the soil (*Rs*=*Pd* - *I*max). The maximum infiltration, *I*max , is formulated as

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

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

In wet season, the upper layer soil is easy to be saturated, resulting in overland flow for a precipitation ($R_s = \max\{P_d - D_{x1}, \quad 0\}$, D_{x1} is soil moisture deficit). The time lag approach is used as watershed regulation to surface runoff and the calculation results are part of stream flow discharge.

2.3.2 Groundwater

Precipitation recharge to groundwater may flow into rivers as baseflow. The rate of the flow Q_g between the stream and the aquifer is calculated from the difference in hydraulic heads in the stream and the adjacent aquifer using following equation (McDonald and Harbaugh, 1988):

$$Q_g = C_{riv}(H_{riv} - h) \tag{10}$$

where Q_g is the flow between the stream and the aquifer, H_{riv} is the head in the stream, 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, recharge and groundwater evapotranspiration depend on groundwater tables, which is described by the governing equation, in 2-dimensional form, as following:

$$\frac{\partial}{\partial x}(Kh\frac{\partial h}{\partial x}) + \frac{\partial}{\partial y}(Kh\frac{\partial h}{\partial y}) = S_y \frac{\partial h}{\partial t} - W_s \tag{11}$$

where S_y is specific yield; and W is a volumetric flux per unit volume representing sources and/or sinks of water, including equation (10), with $W>0$ and $W<0$ for flow in and out of the groundwater system, respectively. The finite-difference groundwater model MODFLOW is used for solving equation (11).

2.4 Water exchanges between unsaturated and saturated zone

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

$$W_e = K(\Psi)\left(\frac{\partial \Psi}{\partial z} - 1\right)_4 = D\left(\frac{\partial \theta}{\partial z}\right)_4 - K_4 \tag{12}$$

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. *We* is the recharge (drainage from the vadose zone) P_{rg} or groundwater loss to the soil and by transpiration E_g.

3 Integration of numerical models

The equations for simulation of soil moisture and groundwater dynamics due to precipitation infiltration and evapotranspiration are coupled to calculation water transfer among precipitation, soil moisture, surface water and groundwater. This couple is based on numerical approaches by discretizing the whole catchment into grids, each of which is hydrological and hydrogeological uniform. Vertically, coupling of soil moisture dynamics and groundwater flow is based on the interface water exchange between unsaturated and saturated zone in equation (12). Surface water and groundwater interaction is based on equation (10). The whole modeling system is shown in Fig 1. Additionally, the model includes artificial influences, e.g. groundwater pumping for irrigation and artificial ponds, to the water exchanges.

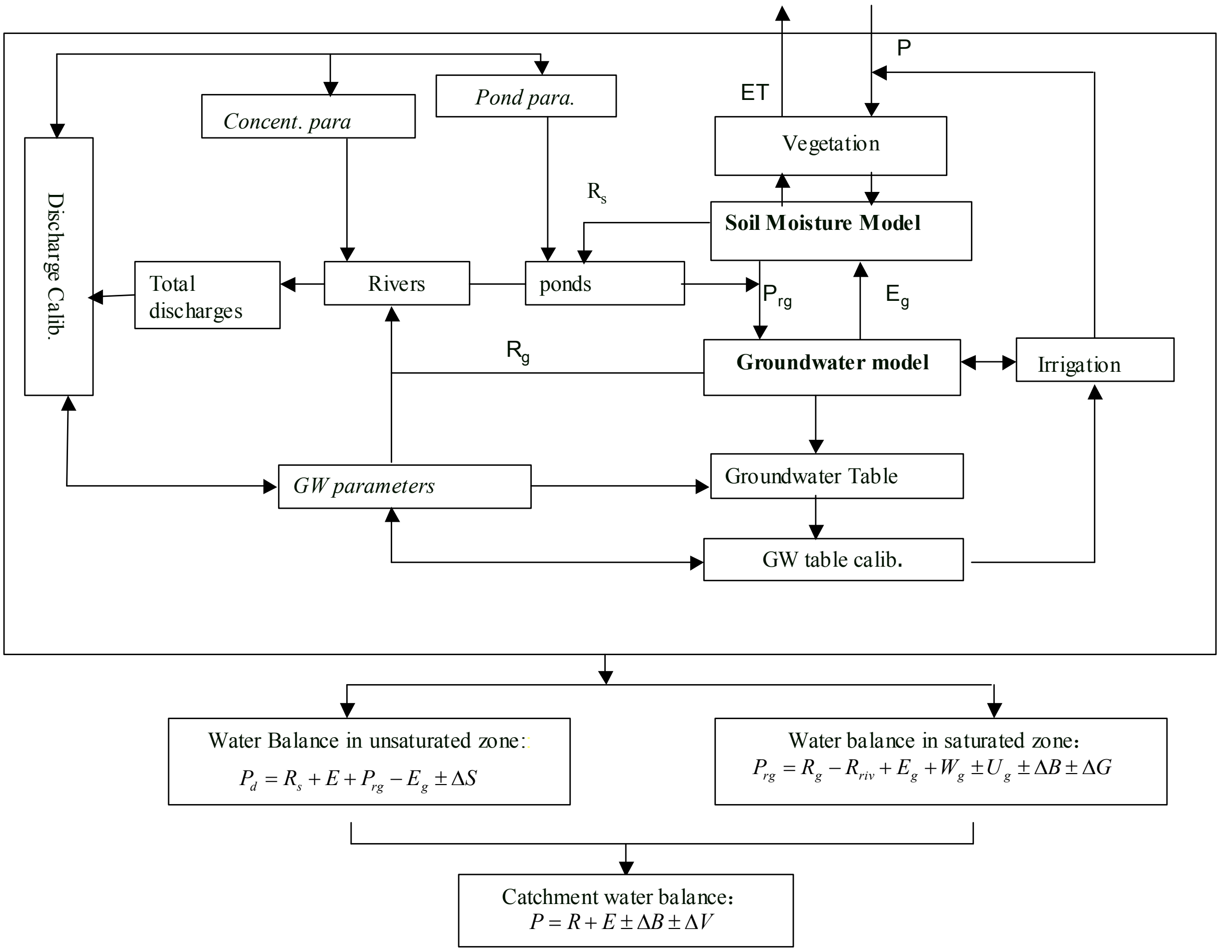


Fig. 1 Scheme of model execution

Model parameters are divided into two types: physical based parameters, e.g. saturated soil moisture content, field capacity, wilting capacity, hydraulic conductivity, specific yield, which are determined primarily regarding soil properties. The left is from model's calibration using streamflow discharges and groundwater tables.

The model is used to calculate streamflow discharges, groundwater tables, soil moisture contents, and infiltration, surface runoff, groundwater recharge and evapotranspiration loss as well.

4 Applications

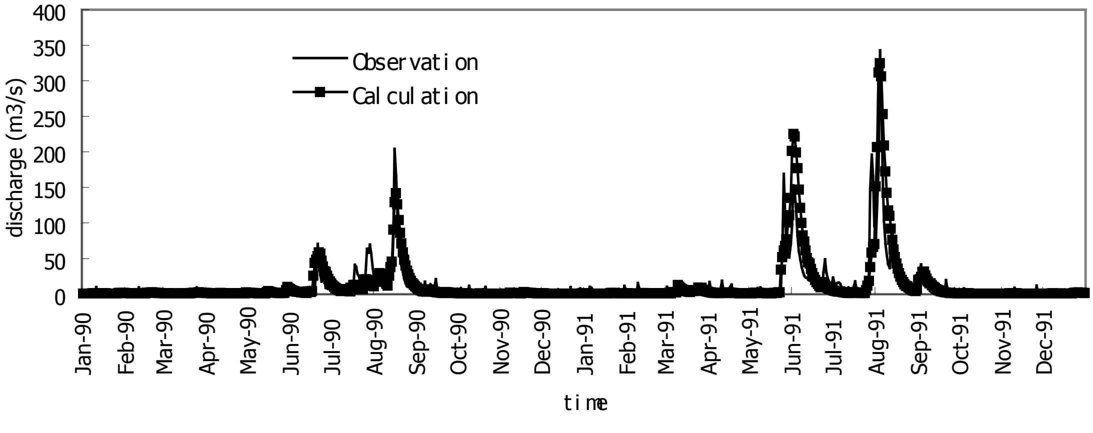
4.1 Model calibration and validation

The model was applied in the Linhuanji catchment of Huaihe plain region, a semi-humid or semi-dry region. Annual precipitation during 1986~1995 is 713mm, approximately 60~70 percent of the precipitation concentrates on summer season of June ~ September. The annual potential evapotranpiration is 960 mm. The catchment area is 2560km^2.

Variations of groundwater tables, similar to the ground surface, are from 45 m in height in the north to 28 m in the south. The depths to groundwater table decrease from 7~8m in the north to 2 m in the south. Annual groundwater variations is approximately 1~2 m.

For estimation of water transfers in the region, we collected ten years data from 1986 to 1995, including daily precipitation of 25 observation stations, pen evaporation, groundwater tables in a five day interval from 30 observation stations, daily streamflow discharge from catchment outlet. Besides, spatial distributions of soil properties and vegetation are available, sandy loam in the north and silt loam in the south. Wheat, maize and sorghum are main crops in the region.

The study region was discretized into 2356 grid units, each 1047 m long and 1048 width. The physical parameters of the soil moisture dynamics are specified by the soil analysis of Cosby et al. (1984), and hydraulic conductivity and specific yield in the saturated zone are 4.4 m/d and 0.055, respectively, in the sand loam region, and 2.8 m/d and 0.045, respectively, in the silt loam regions. The model calculation time interval is one day. The left parameters are calibrated from stream discharges and groundwater tables. Simulated and observed discharges, 1990~1991 and 1994 ~ 1995 as examples, is shown in Fig. 2. Simulated stream discharges usually well represent the observation. Some larger errors between the observed and simulated result from complicated artificially built ponds and dams and irrigation water which are difficult to be estimated accurately. Fig 3 demonstrates that the simulated groundwater tables generally match the observation groundwater tables well.



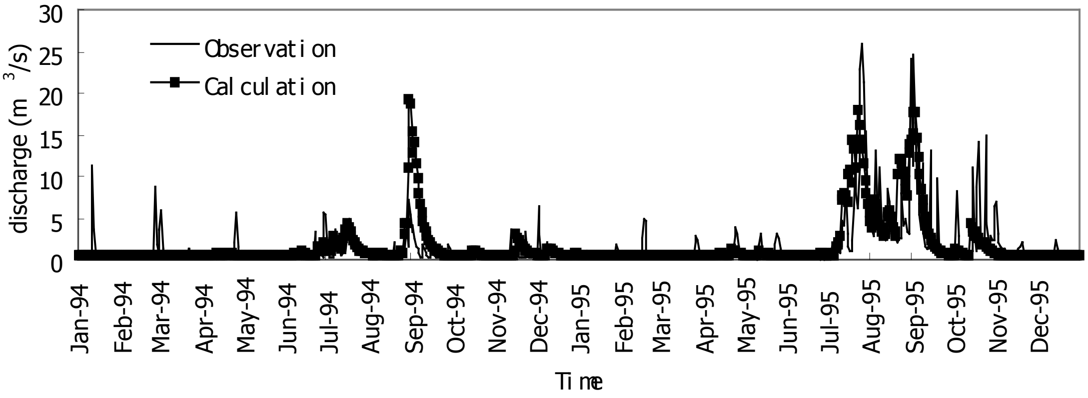
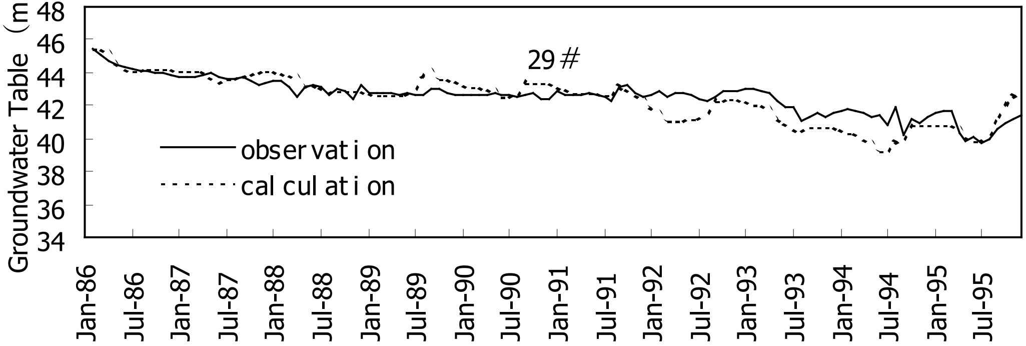
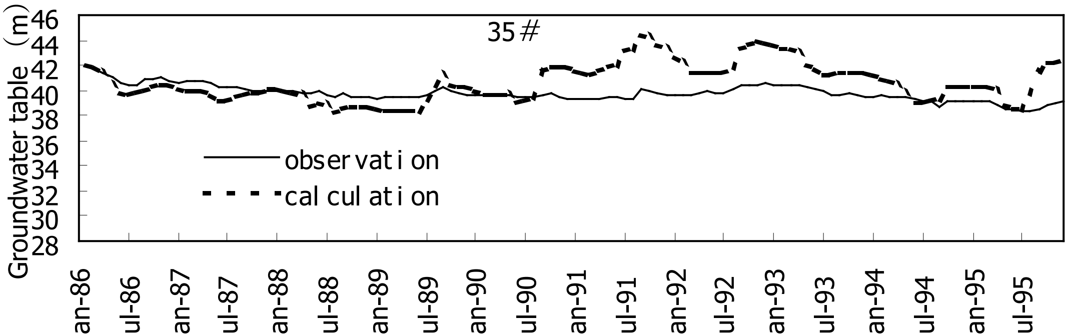
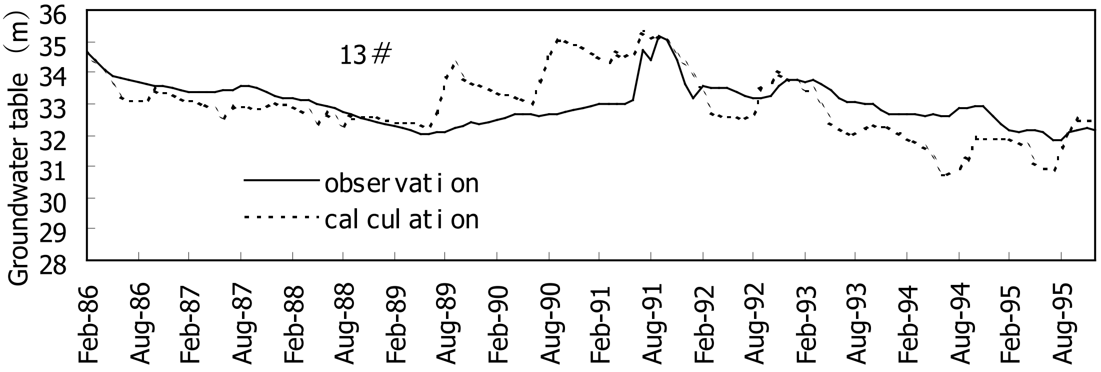


Fig 2 Observed and simulated discharges







Area average

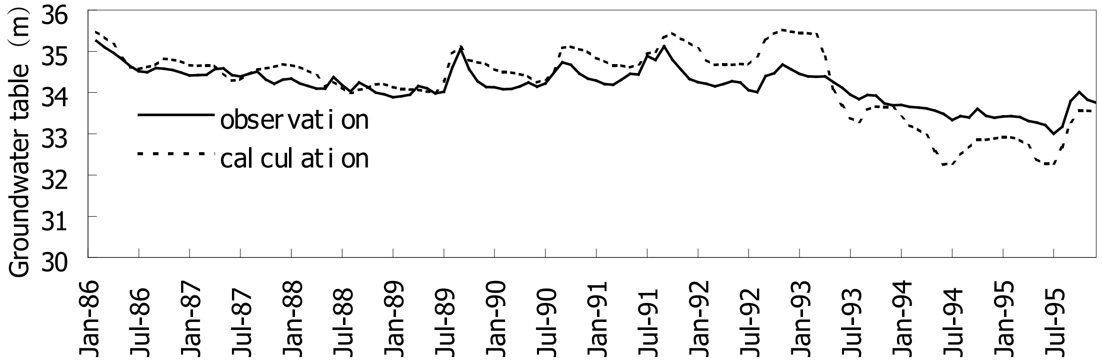


Fig 3 Simulated and observed groundwater tables

4.2 Water budget

Table 1 is calculation results of precipitation transferring into the recharge, surface runoff, total streamflow and baseflow, and losing for evapotranspiration. For the ten-year average, approximately 90 percent of precipitation and irrigation water losses for evapotranspiration; 14.2 percent of precipitation becomes runoff or yearly mean runoff coefficient is 0.24, varying from 0.24 in wet year of 1991 to 0.03 of drought years means for 1986, 1988, 1993 and 1994. 15 percent of precipitation recharges into aquifer or yearly mean recharge coefficient is 0.15, varying from 0.03 to 0.20. Approximately half of the recharge amount losses for evapotranspiration and flows into stream channel as baseflow, and the left remains in the aquifer.

Table 2 lists water balances in unsaturated and saturated zones. For unsaturated zone, the inputs include precipitation P and irrigation water from groundwater withdrawal and upflow 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 one percent. For saturated zone, the recharge from the bottom of unsaturated zone and net inflow from boundary is the aquifer input. Portion of the recharge flows into stream as baseflow or losses through evapotranspiration and withdrawal for irrigation. The water budget in the unsaturated and saturated zones indicates that the model calculations keep water balances well in the study region.

Table 1 water budget and relative errors between observed and calculated runoff

year	P mm	Evapotrans-piration mm	Eg mm	Rs mm	P_{rg} mm	baseflow mm	GW withdrawal for irrigation mm	Total runoff mm	Obs. discharge mm	Relative error %
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 balances in unsaturated and saturated zones

Unsaturated zone					
Input（mm）	P	Eg	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	GW withdrawal	Eg		
	14.64	55.46	45.55		0

5 Conclusions

A comprehensive modeling system based on soil moisture and groundwater dynamics in numerical solutions was developed to simulate hydrological processes of precipitation recharge, surface water, groundwater, and soil moisture content and groundwater tables. The model has been successfully applied in plain area of Linhuanji catchment. Calculations are based on data in detail on meteorology, topography, soil, crops and hydrologic data on stream discharge and groundwater tables. The model is very useful for water resources estimation and planning. The capability of soil moisture content prediction enable the model suitable for scheduling agricultural irrigation planning.

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