Estimation of soil water content and evapotranspiration from irrigated cropland on the North China Plain

Jie Jiang1, Yongqiang Zhang2*, Martin Wegehenkel3, Qiang Yu1, and Jun Xia1

1 Institute of Geographic Sciences and Natural Resources Research, Chinese Academy of Sciences, No Jia 11., Datun Road, Anwai, Beijing 100101, China
2 CSIRO–Land and Water, GPO Box 1666, Canberra ACT 2601, Australia
3 Department of Landscape System Analysis, Centre of Agricultural Landscape and Land use Research (ZALF), Eberswalder Str. 84, 15374 Müncheberg, Germany

Abstract

For nearly 30 y, cropland on the North China Plain (NCP) has been irrigated primarily by pumping groundwater with no sustainable management strategy. This has caused a continuous decline of the water table. A sustainable groundwater management and irrigation strategy must be established in order to prevent further decline of the water table; to do this, one must quantify soil water content and daily rates of deep percolation and locate evapotranspiration from irrigated cropland. For that purpose, we developed a three-layer soil–water balance (SWB) model based on an approach described by Kendy et al. (2003). In this model, the unsaturated soil zone is divided into three layers: a surface active layer, a middle active soil layer, and a lowest passive soil layer. The middle and the lowest layers dynamically change with the development of crop rooting depth. A simple “tipping bucket” routine and an exponential equation are used to redistribute soil water in the three soil layers. The actual evapotranspiration estimated is partitioned into soil evaporation and crop transpiration using a dual crop coefficient reference approach. At first, the model was calibrated using data obtained from five deficiently irrigated field plots located at an experimental site in the NCP between 1998 and 2003. Then, the model was validated by comparing estimated soil water contents with measured ones at three other plots with nondeficient irrigation. The estimates of actual evapotranspiration were compared with those measured by the large-scale weighing lysimeter (3 m²). The index of agreement (IA) for soil water contents varied between 0.62 and 0.80; the concordance correlation coefficient (CCC) and the root mean square error obtained from the same comparison were 0.34–0.65 and 0.043–0.074 cm³ cm⁻³, respectively. The rates of 10 d mean evapotranspiration estimated by the model show a good fit to those measured by the large-scale lysimeter; this is indicated by $IA = 0.94$ and $CCC = 0.88$. Our results indicate that at the irrigated cropland on the plain, deep soil water–percolation rates are usually $<200$ mm y⁻¹ under nondeficient-irrigation conditions.

Key words: soil water balance / soil water content / evapotranspiration / cropland / model / North China Plain

Accepted September 18, 2007

1 Introduction

The North China Plain (NCP), with an area of $3.5 \times 10^5$ km², is one of most important centers of agricultural production in China and home to >200 million people and to China’s capital Beijing, which is located in its northern part. A lack of water limits crop production on the NCP (Zhang et al., 2004). Because of the monsoon, rainfall is spatially and temporally variable. Mean annual precipitation ranges between 400 and 600 mm y⁻¹, of which most falls from June to September. However, mean annual crop evapotranspiration ($ET_a$) under normal irrigation ranges between 800 and 900 mm y⁻¹, which exceeds annual precipitation (Liu et al., 2002). To maintain a high crop-grain yield, crops are irrigated by pumping groundwater. Because agricultural and industrial demand for groundwater has increased rapidly over the last two decades, the water table has dramatically declined in many areas of the plain (Zhang et al., 2001). For policy-makers to scientifically manage the limited water resources in this area, it is crucial to correctly quantify actual crop evapotranspiration ($ET_a$), soil water content, and groundwater recharge in irrigated cropland. In order to quantify these values, an appropriate, precise, physically based, and easy to use modeling approach is required.

Several approaches have been applied to estimate soil water balance (SWB) and $ET_a$ so as to optimize irrigation procedures in the plain (Li et al., 2005; Liu et al., 1998; Shang et al., 2004; Wang et al., 2001; Zhu et al., 2005). These approaches are calibrated and validated across a wide range of water content and irrigation conditions, although long-term...
soil–water content data are not available. Other studies do not consider deep water percolation in the soil–water balance equation (Huang, 2004; Liu et al., 1998). In order to estimate deep water percolation in unsaturated soil layers, a simple mechanistic ten-layer SWB model was developed and applied to irrigated cropland in the plain (Kendy et al., 2003). The data required for the model are readily obtainable; however, the model does not estimate ETg rates well.

In this study, the main objectives were (1) to further simplify this mechanistic ten-layer SWB model, (2) to provide an approach to simulate soil water content at root depth, and (3) to estimate ETg more correctly.

2 Material and methods

2.1 The variable three-layer soil–water balance model

We developed a variable three-layer SWB model that is simpler to operate, in which soil water percolation is based on the model of Kendy et al. (2003). In the new model, the soil reservoir is divided into three soil layers. The lower boundaries of a surface active layer and lowest passive layer are fixed, while that of a middle active layer is movable (Fig. 1). The surface active layer is fixed at a depth of 200 mm; the lowest passive layer is set at the maximum rooting depth, 2000 mm. The depth of the middle active layer dynamically varies according to rooting development (Fig. 1). If the actual crop rooting depth is shallower than the surface active layer, the middle layer disappears, if the crop root extension is deeper than the surface active layer, but does not reach the maximum rooting depth, the three layers are used, and when the roots attain the maximum rooting depth, the lowest soil layer disappears, and drainage from the lower boundary of the passive layer is regarded as deep percolation out of the soil reservoir.

Input data for the three-layer SWB model include daily meteorological data, crop data, initial soil-moisture data, and irrigation data. Meteorological data consist of daily maximum and minimum temperatures (°C), daily mean wind speed (m s–1), daily sunshine duration (h), daily mean relative humidity, and daily precipitation rates (mm d–1). Crop data consist of crop height (m) and rooting depth (mm). The parameters used in the model are listed with their meaning, symbol, unit, status, and method used to obtain them in Tab. 1. Outputs of the model include volumetric soil water content (cm3 cm–3) in the three soil layers, soil water storage in the three soil layers (mm), soil evaporation (mm d–1), crop transpiration (mm d–1), and percolation from the bottom of the passive soil layer (mm d–1).

2.1.2 Soil–water balance equations

The soil–water balance equations are based on the following assumptions. First, infiltration and evapotranspiration are treated as separate, sequential processes. Second, soil evaporation is limited to the upper active layer with a depth of 150 mm. Finally, transpiration is partitioned between the upper layer and the middle layer with an exponential function of layer thickness and rooting depth. Based on these assumptions, the soil–water balance equations in the three soil layers are:

\[ \begin{align*}
\theta_{t,1}L_{t,1} &= \theta_{t-1,1}L_{t-1,1} + h_t + P_t - RO_t - q_t - E_t - T_t, \\
\theta_{t,2}L_{t,2} &= \theta_{t-1,2}L_{t-1,2} + q_{t-1,1} - q_{t,2} + \Delta R_t \theta_{t-1,3} - T_{t,2}, \\
\theta_{t,3}L_{t,3} &= \theta_{t-1,3}L_{t-1,3} + q_{t,2} - q_{t,3},
\end{align*} \]

where \( t \) is the Julian date, 1, 2, and 3 are the soil layers from the surface to the maximum root depth (2000 mm), respectively, \( \theta \) is the volumetric soil moisture (cm3 cm–3), \( L \) is the thickness of the soil layer (mm), \( P \) is the precipitation (mm d–1), \( I \) is the applied irrigation (mm d–1), \( RO \) is the surface runoff (mm d–1), \( E \) is the soil evaporation (mm d–1), \( T \) is the crop transpiration (mm d–1), \( q \) is the downflow flux (mm d–1), and

![Figure 1: Sketch map of soil–water balance components in the three-layer soil–water balance model](image)

* 1: surface active soil layer; 2: middle active soil layer; 3: bottom passive soil layer.

\( P \) = precipitation; \( I \) = irrigation; \( RO \) = surface runoff; \( T \) = crop transpiration; \( E \) = soil evaporation; \( \Delta R \) = daily increase of rooting depth; \( q_1 \) = downflow from the bottom of the upper layer; \( q_2 \) = downflow from the bottom of the middle layer; \( q_3 \) = downflow from the bottom of the lowest layer.
\( \Delta R \), which equals \( L_{2} - L_{1,2} \), is the rate of increase of crop rooting depth (mm d\(^{-1}\)).

### 2.1.3 Percolation submodel

The percolation submodel according to Kendy et al. (2003) estimates saturated and unsaturated water fluxes in the soil layers and assumes that gravity forces dominate over matrix forces in the soil. Modeled flux is therefore always downward. First, precipitation or irrigation is added to the upper soil layer, which is filled to saturation. Water in excess of the layer’s saturated soil moisture is distributed downward to the deeper layers in a “tipping bucket” fashion until either each layer is filled to saturation or all of the water has been distributed. Excess water that drains from the bottom of the passive soil layer is treated as groundwater recharge.

Assuming no incoming or outgoing water flux other than that produced by the unit gradient at the bottom of the layer, outflow from a layer can be expressed according to the conservation of mass as

\[
\frac{d\theta}{dt} = -K(\theta),
\]

where \( K \) is the unsaturated hydraulic conductivity (mm d\(^{-1}\)). Equation 4 is solved using the following exponential relationship between \( K \) and soil water content (Kendy et al., 2003):

\[
K(\theta) = K_{sat} \exp\left(-a\frac{\theta_s - \theta}{\theta_s - \theta_d}\right)
\]

where \( K_{sat} \) is the saturated hydraulic conductivity, \( \theta_s \) is the volumetric moisture content of the soil layer at saturation, \( \theta_d \) is the soil water content at wilting point, and \( a \) is a site-specific parameter determined mainly from physical soil characteristics. For the same soil–water content conditions, the decrease of \( K \) is steeper for greater values of \( a \) (Fig. 2).

By substituting Eq. 4 into Eq. 5, separating variables, and integrating Eq. 5, we obtain the volumetric soil water content of a single layer after percolation (Zhang and Wegehenkel, 2006):

\[
\theta_t = \theta_s - \frac{\theta_s - \theta_d}{a} \ln\left[ \frac{a \cdot K_{sat} \Delta t}{L(\theta_s - \theta_d)} + \exp\left( \frac{a(\theta_s - \theta_{t-\Delta t})}{\theta_s - \theta_d} \right) \right]
\]

The flux into the deeper layer is the difference between \( \theta_t \) and \( \theta_{t-\Delta t} \).

### 2.1.4 Crop evapotranspiration submodel

The dual crop coefficient method developed by the Food and Agriculture Organization (FAO) of the United Nations was used to estimate crop \( ETa(Allen et al., 1998) \). Estimating crop \( ETa \) first requires obtaining a reference evapotranspiration \( (ET0 [mm d\(^{-1}\)]) \), which is computed from weather data using the FAO-Penman-Monteith equation:

\[
ET0 = \frac{0.408 \Delta (R_n - G) + \gamma \left( \frac{900}{T_a + 273} \right) \Delta (e_s - e_a)}{\Delta + \gamma (1 + 0.34 \Delta b_s)},
\]

where \( R_n \) is the net radiation at the crop surface (MJ m\(^{-2}\) d\(^{-1}\)), \( G \) is the soil heat flux density (MJ m\(^{-2}\) d\(^{-1}\)), \( T_a \) is the daily mean air temperature at 2 m height (°C), \( u_2 \) is the wind speed at 2 m height (m s\(^{-1}\)), \( e_s \) is the saturation vapor pressure (kPa), \( e_a \) is the actual vapor pressure (both in kPa) \( (e_a = e_s \times RH, RH \ is \ relative \ humidity) \), \( \Delta \) is the slope of the vapor-pressure curve (kPa °C\(^{-1}\)), and \( \gamma \) is a psychrometric constant (kPa °C\(^{-1}\)). \( R_n \) is the difference between incoming net short-wave radiation \( (R_{ns}) \) and outgoing net long-wave radiation \( (R_{ls}) \). \( R_{ns} \) is estimated from surface solar radiation \( (R_s) \):

\[
R_{ns} = (1 - \alpha_s)R_s
\]

where \( \alpha_s \) is the albedo of the reference grassland, alfalfa. \( R_s \) is estimated as

\[
R_s = (a_d + b_d \cdot \frac{S}{N} \cdot R_a),
\]

where \( S \) is actual sunshine duration (h), \( N \) is maximum possible sunshine duration (h), \( S/N \) is relative sunshine duration \( (0–1) \), \( R_a \) is extraterrestrial radiation (MJ m\(^{-2}\) d\(^{-1}\)), and \( a_d \) and \( b_d \) are empirical coefficients. Soil heat flux density \( G \) is estimated with a simple calculation based on the idea that soil temperature follows air temperature. The detailed algorithms for \( R_n \) and \( G \) can be obtained from the FAO 56 ET method (Allen et al., 1998).

Actual evapotranspiration \( ETa \) is determined by using \( ET0 \) combined with the basal crop coefficient \( (K_{eb}) \), soil evaporation \( (K_e) \), and water stress \( (K_s) \):

\[
ETa = (K_{eb} K_s + K_e) ET0.
\]

According to Eq. 10, \( ETa \) is partitioned into actual transpiration \( (T_{tot}) \) and soil evaporation \( (E) \):

\[
T_{tot} = K_{eb} K_s ET0 \quad \text{and} \quad E = K_e ET0.
\]
The basal crop coefficient $K_{cb}$ can be estimated from a given basal crop coefficient ($K_{cb0}$), meteorological data, and crop height (Allen et al., 1998):

$$K_{cb} = K_{cb0} + 0.04(u_2 - 2) - 0.004([RH_{min} - 0.45])$$

$$\times \left( \frac{h_c}{3} \right)^{0.3}$$

where $RH_{min}$ is the mean value for minimum relative humidity during the mid- or late-season growth stage for $0.2 \leq RH_{min} \leq 0.8$, and $h_c$ is crop height (m). $K_{cb0}$ varies for the initial, middle, and end crop phenological stages (Tab. 1) and is determined mainly by crop type, while $K_{cb0}$ is influenced by local conditions, cultural practices, and crop variety as well. In most cases, local values for $K_{cb0}$ do not deviate by $>0.2$ from the values recommended by the FAO 56 ET method (Allen et al., 1998).

The water-stress coefficient $K_s$, which varies from 0 to 1, is a dimensionless transpiration-reduction factor defined as follows:

$$K_s = \frac{TAW - Dr}{TAW - RAW} = \frac{TAW - Dr}{(1 - p)TAW}$$

where $Dr$ is root-zone depletion (mm), $TAW$ is the total available soil water in the root zone (mm) ($TAW = (\theta_f - \theta_d)L_r$; $\theta_f$ is

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Meaning</th>
<th>Unit</th>
<th>Used in</th>
<th>Way of obtaining</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$K_{sat}$</td>
<td>saturated hydraulic conductivity</td>
<td>mm d$^{-1}$</td>
<td>Eqs. 5 and 6</td>
<td>optimized</td>
<td>500 for layer 1; 100 for layers 2 and 3</td>
</tr>
<tr>
<td>$\theta_s$</td>
<td>saturated soil water content</td>
<td>cm$^3$ cm$^{-3}$</td>
<td>Eqs. 5 and 6</td>
<td>measurements (Zhang et al., 2002)</td>
<td>0.433 for layer 1; average at the depths of layers 2 and 3</td>
</tr>
<tr>
<td>$\theta_f$</td>
<td>field capacity</td>
<td>cm$^3$ cm$^{-3}$</td>
<td>calculating TAW and TEW</td>
<td>measurements (Zhang et al., 2002)</td>
<td>0.364 for layer 1; average at the depths of layers 2 and 3</td>
</tr>
<tr>
<td>$\theta_d$</td>
<td>soil water content at wilting point</td>
<td>cm$^3$ cm$^{-3}$</td>
<td>Eqs. 5 and 6</td>
<td>measurements (Zhang et al., 2002)</td>
<td>0.096 for layer 1; average at the depths of layers 2 and 3</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>exponential coefficient in hydraulic-conductivity equation</td>
<td>–</td>
<td>Eqs. 5 and 6</td>
<td>optimized</td>
<td>22 for layer 1; 15 for layers 2 and 3</td>
</tr>
<tr>
<td>$\alpha_c$</td>
<td>albedo for the reference grassland</td>
<td>–</td>
<td>Eq. 8</td>
<td>Allen et al. (1998)</td>
<td>0.23</td>
</tr>
<tr>
<td>$a_x, b_x$</td>
<td>empirical coefficients in solar-radiation estimations</td>
<td>–</td>
<td>Eq. 9</td>
<td>optimized according to measurements</td>
<td>0.29 and 0.44</td>
</tr>
<tr>
<td>$K_s$</td>
<td>water-stress coefficient</td>
<td>–</td>
<td>Eqs. 10 and 11</td>
<td>defined in Eq. 13</td>
<td>variable</td>
</tr>
<tr>
<td>$K_{eo}$</td>
<td>coefficient for soil evaporation</td>
<td>–</td>
<td>Eqs. 10 and 11</td>
<td>defined in Eq. 17</td>
<td>variable</td>
</tr>
<tr>
<td>$K_{cb}$</td>
<td>basal crop coefficient</td>
<td>–</td>
<td>Eq. 10</td>
<td>defined in Eq. 12</td>
<td>variable</td>
</tr>
<tr>
<td>$K_{cb0}$</td>
<td>given basal crop coefficient</td>
<td>–</td>
<td>Eq. 12</td>
<td>Allen et al. (1998)</td>
<td>0.15, 1.1, and 0.3 for initial, middle, and end stage, respectively</td>
</tr>
<tr>
<td>TAW</td>
<td>total available soil water in the root zone</td>
<td>mm</td>
<td>Eq. 13</td>
<td>$(\theta_f - \theta_d)L_r$</td>
<td>variable according to $L_r$</td>
</tr>
<tr>
<td>RAW</td>
<td>readily available soil water in the root zone</td>
<td>mm</td>
<td>Eq. 13</td>
<td>$pTAW$</td>
<td>variable</td>
</tr>
<tr>
<td>$Dr$</td>
<td>root-zone depletion</td>
<td>mm</td>
<td>Eq. 13</td>
<td>$\theta_f - 0.5\theta_dL_r$</td>
<td>variable</td>
</tr>
<tr>
<td>$p$</td>
<td>average fraction of TAW</td>
<td>–</td>
<td>Eq. 13</td>
<td>Eq. 14</td>
<td>variable</td>
</tr>
<tr>
<td>$p_{give}$</td>
<td>given threshold for $p$</td>
<td>–</td>
<td>Eq. 14</td>
<td>set constant</td>
<td>0.55</td>
</tr>
<tr>
<td>$u_f$</td>
<td>transpiration-uptake fraction</td>
<td>–</td>
<td>Eq. 15</td>
<td>defined in Eq. 16</td>
<td>variable according to $L_r$</td>
</tr>
<tr>
<td>$w_p$</td>
<td>water use–distribution coefficient</td>
<td>–</td>
<td>Eq. 16</td>
<td>Novak, 1987</td>
<td>3.64</td>
</tr>
<tr>
<td>$K_r$</td>
<td>dimensionless evaporation-reduction coefficient</td>
<td>–</td>
<td>Eq. 17</td>
<td>defined in Eq. 18</td>
<td>variable</td>
</tr>
<tr>
<td>$K_{max}$</td>
<td>maximum crop coefficient</td>
<td>–</td>
<td>Eq. 17</td>
<td>Eq. 72 (Allen et al., 1998)</td>
<td>1.05–1.30</td>
</tr>
<tr>
<td>$l_{ew}$</td>
<td>fraction of the soil that is both exposed and wetted</td>
<td>–</td>
<td>Eq. 17</td>
<td>Eq. 75 (Allen et al., 1998)</td>
<td>variable</td>
</tr>
<tr>
<td>TEW</td>
<td>maximum cumulative depth of soil evaporation</td>
<td>mm</td>
<td>Eq. 18</td>
<td>$(\theta_f - 0.5\theta_d)L_r$</td>
<td>47</td>
</tr>
<tr>
<td>REW</td>
<td>readily evaporable water in surface soil layer</td>
<td>mm</td>
<td>Eq. 18</td>
<td>set constant</td>
<td>15</td>
</tr>
</tbody>
</table>
the field capacity, \( L_r \) is rooting depth), \( RAW \) is the readily available soil water in the root zone (mm) \( (RAW = p TAW) \), and \( p \) is the average fraction of \( TAW \) that can be depleted from the root zone before moisture stress occurs. Root-zone depletion \( D_r \) is calculated from the soil–water balance equation in the root zone (see Eq. 85 in Allen et al. [1998]). Values of \( p \) are higher at low rates of \( ET_a \) under non-drought-stressed conditions than at high rates. An empirical approximation for adjusting \( p \) for \( ET_a \) rate is

\[
p = p_{\text{give}} + 0.04 (5 - ET_a), \tag{14}
\]

where \( p_{\text{give}} \) is a given threshold used as a constant for \( p \), limited to 0.1–0.8; \( p_{\text{give}} \) differs from one crop to another. For cereals such as barley, wheat, and maize, it has a value of 0.55 for \( ET_a = 5.0 \text{ mm d}^{-1} \) (Allen et al., 1998). \( T_{tot} \) is partitioned into transpiration from the first layer \( (T_1) \) and that from the second layer \( (T_2) \) according to a transpiration uptake fraction \( (\phi) \):

\[
T_1 = (1 - \phi) T_{tot} \quad \text{and} \quad T_2 = \phi T_{tot}. \tag{15}
\]

When \( L_r < L_t \), \( \phi \) equals 0. When \( L_r > L_t \), \( \phi \) is estimated as follows, according to the assumed exponential distribution of crop roots:

\[
\phi = \left( \frac{1}{1 - \exp(-w_u)} \right) \left\{ \exp\left[ -w_u \left( \frac{L_t}{L_r} \right) \right] \right. \\
\left. \times \left[ 1 - \exp(-w_u \left( \frac{L_t - L_r}{L_r} \right) ) \right] \right\}, \tag{16}
\]

where \( w_u \) is a water-use distribution coefficient, which varies between 0 and 5 (Kendy et al., 2003); \( w_u \) is usually obtained from optimization or expert knowledge.

Soil evaporation \( K_e \) is calculated from

\[
K_e = \min[K_r (K_{\text{max}} - K_{\text{cd}}), f_{\text{ew}} K_{\text{max}}], \tag{17}
\]

where \( K_r \) is the dimensionless evaporation-reduction coefficient depending on the cumulative depth of water evaporated from the topsoil, \( f_{\text{ew}} \) is the fraction of soil that is both exposed and wetted, i.e., the fraction of soil surface from which most evaporation occurs, and \( K_{\text{max}} \) is the maximum value of the crop coefficient. \( K_r \) is calculated from

\[
K_{r,t} = \begin{cases} 
1 & \text{if } D_e \leq \text{REW} \\
\frac{T_{EW} - D_{e,t-\Delta t}}{T_{EW} - \text{REW}} & \text{if } D_e > \text{REW}
\end{cases}, \tag{18}
\]

where \( D_{e,t-\Delta t} \) is the cumulative depth of evaporation in mm from the surface layer at the end of day \( t - \Delta t \), \( T_{EW} \) is the maximum cumulative depth of evaporation from the soil surface layer, and \( \text{REW} \) is the readily evaporable water, i.e., the maximum depth of water that can be evaporated from the topsoil layer without any restrictions (Tab. 1).

2.2.2 Experimental data

The physical parameters of the soil profile, including saturated soil water content, field capacity, water content at the wilting point, and bulk density, were measured by sampling undisturbed soil columns at the seven soil depths 0–200, 200–350, 350–650, 650–950, 950–1450, 1450–1700, and 1700–1900 mm (Zhang et al., 2002). The parameter values of the 0–200 mm soil layer were used directly as the parameters for the first layer of the soil–water balance model, while those in lower soil layers were averaged to calculate corresponding parameters for the second and third layers of the model.

The daily meteorological data (including maximum temperature, minimum temperature, wind speed, relative humidity, and sunshine hours) were measured by an automatic meteorological observation station where grassland was cropped and normally irrigated during the experimental period 1998 to 2003 (Changchun Meteorological Instrument Research Institute, China). Daily precipitation rates were measured by summing up hourly tipping-bucket measurements. The rooting depths of winter wheat and maize were reported by Zhang (1999) as a function of the number of days since planting.

The \( ET_a \) has been measured by a large-scale weighing lysimeter since 1995. The lysimeter was constructed by The Institute of Geographical Science and Natural Resources Research, a part of the Chinese Academy of Sciences. It is 3 m² in area and 2.5 m deep and contains approx. 14,000 kg of undisturbed soil. It can accurately measure daily actual \( ET \) with 0.02 mm precision. The lysimeter has four parts: the main body, the weighing system, the drainage system, and the data-collection system. It is described in detail by Zhang et al. (2002). In the data-collection system, a signal amplifier
detects changes of displacement and pressure. Because the numerical data collector malfunctioned, data from the signal amplifier were read manually at 8:00 a.m. and 8:00 p.m., and the difference of the observation results was used to estimate daily ET rates. The lysimeter, which was surrounded by cropland, was situated approx. 20 m away from experimental field plots. Double-cropped winter wheat and summer maize, the same as on the adjacent field plots, were planted at the lysimeter, and irrigation for the lysimeter was the same as for the field plots with nondeficient-irrigation application. This means that the soil water content at the rooting depth was >80% field capacity.

The 16 field plots, each with an area of 50 m², were used for agricultural water-management experimentation since 1998 (Zhang et al., 2004). Each site was well irrigated prior to the model-calibration period. Five different irrigation schedules were applied for winter wheat and two were applied for maize (Tab. 2). The five plots 1, 2, 3, 4, and 7 were randomly irrigated with treatments A to E during 1998 to 2003; these were used for model calibration. Plots 5, 6, and 8 were selected for model validation. Among them, plot 8 was irrigated with treatment A during the whole experimental period; plots 5 and 6 were irrigated with treatment E in the spring of 1998. Neutron probes (Institute of Hydrology, UK) were used for soil-water measurements at 20 cm intervals at depths of 0–200 cm. The calibration equation at depths of 140–200 cm was assumed identical to that at depths 120–140 cm. Gravimetric samples for the soil depth 0–20 cm were taken simultaneously to calibrate neutron-probe measurements.

Net radiation above the cropland was measured in 1999 with a Q7–1 type net radiometer (Campbell Scientific, Logan, UT) with hemispherical polyethylene domes at Luancheng Station (Zhang et al., 2002). Readings were taken every 10 s, averaged over 20 min intervals, and logged with a data logger (Model CR10X, Campbell Scientific, Logan, UT).

### 2.3 Model calibration

Parameters \( K_{sat} \) and \( \alpha \) were optimized with a nonlinear least-square Gauss-Newton method. The cost function was the lowest root mean square error between measured and estimated soil water content. The model ran for the first 2 years (1998–1999) to warm up. It was then calibrated using the measured soil water content of each fifth day and the simulated one at the same time from 2000 to 2003. Five calibrated plots 1, 2, 3, 4, and 7 were selected to optimize the parameters \( K_{sat} \) and \( \alpha \). We assume that \( K_{sat} \) and \( \alpha \) in the middle and lowest passive layers are identical due to the dynamic change of thickness of these two layers. Therefore, \( K_{sat} \) and \( \alpha \) were optimized in two layers. Averaged \( K_{sat} \) in each layer was obtained from the arithmetic average of the five optimized values for \( K_{sat} \) as well as for averaged \( \alpha \). The average \( K_{sat} \) and \( \alpha \) were then used for simulations in plots 5, 6, and 8.

### 2.4 Validation criteria of diagnostic variables

Model estimates were evaluated with four different approaches: the difference between observed and estimated means (MAD), the index of agreement (IA) (Willmott, 1981, 1984), the concordance correlation coefficient (CCC) (Lin et al., 2002; Lin, 1989), and the root mean square error (RMSE), as follows:

\[
MAD = \mu_1 - \mu_2,
\]

\[
RMSE = \sqrt{\frac{\sum_{i=1}^{N} (Y_1 - Y_2)^2}{N - 1}},
\]

\[
IA = 1 - \frac{\sum_{i=1}^{N} (Y_1 - Y_2)^2}{\sum_{i=1}^{N} \left[ |Y_1 - \mu_2| + |Y_2 - \mu_2| \right]^2},
\]

\[
CCC = \frac{2\sigma_{1,2}}{\sigma_1^2 + \sigma_2^2 + (\mu_1 - \mu_2)^2},
\]

where \( Y_1 \) and \( Y_2 \) are the estimated and measured values, respectively; \( \mu_1 \) and \( \mu_2 \) are the means of estimated and measured values, respectively; \( N \) is the sample number, and \( \sigma_{1,2} \), \( \sigma_1^2 \), and \( \sigma_2^2 \) are vectors in the covariance matrix of estimated and measured pairs. The IA results in a range between 0 and 1; the closer it is to 1, the better the fit between observed and estimated values. The CCC, which evaluates the degree to which pairs fall on the 45° line through the origin, is scaled to between −1 and 1.

---

Table 2: Irrigation treatments for winter wheat and maize at Luancheng Station (1998–2003).

<table>
<thead>
<tr>
<th>Crop</th>
<th>Treatment</th>
<th>Growth stage and irrigation treatment (days)</th>
<th>a</th>
<th>b</th>
</tr>
</thead>
<tbody>
<tr>
<td>Winter wheat</td>
<td>A</td>
<td>winter dormancy: 1.0</td>
<td>a</td>
<td>-</td>
</tr>
<tr>
<td></td>
<td>B</td>
<td>spring green up: 0.8</td>
<td>-</td>
<td>0.8</td>
</tr>
<tr>
<td></td>
<td>C</td>
<td>stem extension: 0.8</td>
<td>0.8</td>
<td>0.8</td>
</tr>
<tr>
<td></td>
<td>D</td>
<td>heading grain filling: 1.0</td>
<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td>E</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Maize</td>
<td>D</td>
<td>-</td>
<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td></td>
<td>E</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>

a \( \theta \) is average soil water content of crop root depth; \( \theta_i \) is average field capacity of crop root depth. 1.0 or 0.8 means the ratio of \( \theta \) to \( \theta_i \).
b "-" shows that no irrigation is applied.
3 Results

3.1 Optimization of parameters

For the upper layer, the optimized $K_{sat}$ varied between 479 and 528 mm d$^{-1}$ in the five calibration plots 1, 2, 3, 4, and 7, with an average $K_{sat}$ of 502 mm d$^{-1}$ and a $K_{sat}$ standard deviation of 17.2 mm d$^{-1}$. For the middle layer as well as in the lowest layer, the optimized $K_{sat}$ changed from 105 to 119 mm d$^{-1}$, with an average value of 113.4 mm d$^{-1}$ and a standard deviation of 6.3 mm d$^{-1}$. For an undisturbed loam soil on the NCP, $K_{sat}$ was measured approx. 2,300 mm d$^{-1}$ for the soil layer from 0 to 250 mm deep and 17–500 mm d$^{-1}$ for deeper soil layers (Deng et al., 2002). The optimized $K_{sat}$ in the upper active layer was evidently lower than the measured one because the surface soil layer at our site was highly disturbed due to continuous cropping (Deng et al., 2002).

The optimized $\alpha$ for the surface layer varied between 15.0 and 21, with a small standard deviation of 2.7 and an average value of 18.5; that for the other two layers varied between 14 and 19, with an average value of 15.5 and a smaller standard deviation of 2.0. The $\alpha$ optimized for the surface layer was slightly higher than that by the trial-and-error adjustment calibration, for which $\alpha$ was estimated at 13–16 in ten soil layers from 0 to 2000 mm depth (Kendy et al., 2003). The optimized $\alpha$ for the other two layers, however, was almost same as that in Kendy et al. (2003).

The calibrated model simulated soil water content in the lower two layers better than in the upper layer. In the upper layer, the cost function RMSE varied between 0.07 and 0.09, with an average of 0.082 and a standard deviation of 0.006. In the lower layers, RMSE changed from 0.04 to 0.07, with an average of 0.057 and a standard deviation of 0.007.

Figure 3: Seasonal dynamics of measured and estimated soil water content (cm$^3$ cm$^{-3}$) at the three soil layers (surface active layer, middle active layer, and lowest passive layer), and evapotranspiration, deep soil-water percolation, precipitation, and irrigation at plot 8 with nondeficient irrigation.
3.2 Model validation: Prediction of soil water content

Figure 3 shows the seasonal variation of measured and simulated soil water content at plot 8. The simulated soil water content corresponded well with that measured, indicated by a RMSE between 0.043 and 0.074 cm$^3$ cm$^{-3}$ and an IA varying between 0.67 and 0.80 in the three soil layers. Simulation quality increased continuously from the upper layer to the middle layer and to the lowest layer.

The simulation quality was further evaluated at plots 5 and 6 (Tab. 4). Comparisons of measured and simulated soil water content at the three plots indicated a MAD between –0.026 and 0.014 cm$^3$ cm$^{-3}$, RMSE between 0.043 and 0.070 cm$^3$ cm$^{-3}$, IA between 0.62 and 0.80, and CCC between 0.34 and 0.65. The middle layer was better simulated than was the upper layer when evaluated by MAD and RMSE but not by IA and CCC. The lowest layer was simulated best compared to the other two layers. The best simulation quality was found at plot 8.

3.3 Model validation: Prediction of evapotranspiration

The seasonal variation of estimated $ET_a$ for the validation plots 5, 6, and 8 agreed well with the $ET_a$ measured by the large-scale lysimeter during 1998–2003 (Fig. 4). Estimated and measured $ET_a$ both showed two peaks in the winter wheat season and the maize season, respectively. Measured and estimated minimum $ET_a$ both appeared during winter dormancy in the winter wheat season. The quality of the estimated $ET_a$ was analyzed by comparing mean measured and estimated $ET_a$ rates over 10 d (Fig. 5). The values of $IA = 0.94$ and $CCC = 0.88$ suggested a good estimate of $ET_a$. A MAD of –0.34 mm d$^{-1}$, however, shows that estimated $ET_a$ tended to be lower than measured rates (Fig. 5).

3.4 Annual outputs of the model

The annual rates of the estimated water-balance components for the three irrigated plots in the period 1998–2003 are summarized in Tab. 3. One year in Tab. 3 is defined from October 1 to September 30 the following year, because in the research area, winter wheat is sown in early October and harvested in mid-June and summer maize is sown in mid-June and harvested in late September. The water surplus due to irrigation and precipitation is mainly transpired by crops and evaporated from soil. At irrigated plots with nondeficient irrigation, estimated annual rates of $ET_a$ were within 800–850 mm y$^{-1}$ during 1998–2002. In the period 2002–2003, the annual rates of $ET_a$ for all the plots were only

Table 3: Estimated soil–water balance components at three irrigated plots from 1998 to 2003. $P$ = Precipitation; $I$ = irrigation; $T_{tot}$ = crop transpiration; $E$ = soil evaporation; $ET_a$ = actual evapotranspiration; $q_3$ = soil-water percolation beneath the maximum rooting depth; $S$ = soil-moisture change.

<table>
<thead>
<tr>
<th>plot</th>
<th>Year</th>
<th>$P$</th>
<th>$I$</th>
<th>$ET_0$</th>
<th>$T_{tot}$</th>
<th>$E$</th>
<th>$ET_a$</th>
<th>$q_3$</th>
<th>$S$</th>
</tr>
</thead>
<tbody>
<tr>
<td>No.5</td>
<td>1988-10-01–1999-09-30</td>
<td>347</td>
<td>487</td>
<td>938</td>
<td>644</td>
<td>159</td>
<td>803</td>
<td>89</td>
<td>–58</td>
</tr>
<tr>
<td></td>
<td>1999-10-01–2000-09-30</td>
<td>402</td>
<td>504</td>
<td>935</td>
<td>684</td>
<td>146</td>
<td>830</td>
<td>95</td>
<td>–19</td>
</tr>
<tr>
<td></td>
<td>2000-10-01–2001-09-30</td>
<td>344</td>
<td>436</td>
<td>937</td>
<td>661</td>
<td>178</td>
<td>838</td>
<td>7</td>
<td>–66</td>
</tr>
<tr>
<td></td>
<td>2001-10-01–2002-09-30</td>
<td>399</td>
<td>160</td>
<td>949</td>
<td>563</td>
<td>170</td>
<td>732</td>
<td>1</td>
<td>–175</td>
</tr>
<tr>
<td></td>
<td>2002-10-01–2003-09-30</td>
<td>476</td>
<td>393</td>
<td>877</td>
<td>579</td>
<td>172</td>
<td>751</td>
<td>0</td>
<td>118</td>
</tr>
<tr>
<td>No.6</td>
<td>1998-10-01–1999-09-30</td>
<td>347</td>
<td>487</td>
<td>938</td>
<td>644</td>
<td>163</td>
<td>807</td>
<td>64</td>
<td>–37</td>
</tr>
<tr>
<td></td>
<td>1999-10-01–2000-09-30</td>
<td>402</td>
<td>553</td>
<td>935</td>
<td>684</td>
<td>146</td>
<td>830</td>
<td>136</td>
<td>–11</td>
</tr>
<tr>
<td></td>
<td>2001-10-01–2002-09-30</td>
<td>399</td>
<td>160</td>
<td>949</td>
<td>597</td>
<td>162</td>
<td>760</td>
<td>6</td>
<td>–206</td>
</tr>
<tr>
<td></td>
<td>2002-10-01–2003-09-30</td>
<td>476</td>
<td>468</td>
<td>877</td>
<td>589</td>
<td>172</td>
<td>761</td>
<td>1</td>
<td>182</td>
</tr>
<tr>
<td>No.8</td>
<td>1998-10-01–1999-09-30</td>
<td>347</td>
<td>473</td>
<td>938</td>
<td>644</td>
<td>156</td>
<td>800</td>
<td>31</td>
<td>–11</td>
</tr>
<tr>
<td></td>
<td>1999-10-01–2000-09-30</td>
<td>402</td>
<td>534</td>
<td>935</td>
<td>684</td>
<td>146</td>
<td>830</td>
<td>101</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>2001-10-01–2002-09-30</td>
<td>399</td>
<td>495</td>
<td>949</td>
<td>656</td>
<td>187</td>
<td>843</td>
<td>1</td>
<td>50</td>
</tr>
<tr>
<td></td>
<td>2002-10-01–2003-09-30</td>
<td>476</td>
<td>680</td>
<td>877</td>
<td>590</td>
<td>172</td>
<td>761</td>
<td>181</td>
<td>213</td>
</tr>
</tbody>
</table>

Table 4: Validation-criteria analysis of soil water content at the three plots 5, 6, and 8. MAD and RMSE are in the same unit of cm$^3$ cm$^{-3}$; $n$ is sample number.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Plot</th>
<th>MAD</th>
<th>RMSE</th>
<th>$IA$</th>
<th>CCC</th>
<th>$n$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>5</td>
<td>–0.026</td>
<td>0.070</td>
<td>0.73</td>
<td>0.53</td>
<td>258</td>
</tr>
<tr>
<td>1</td>
<td>6</td>
<td>–0.026</td>
<td>0.074</td>
<td>0.70</td>
<td>0.49</td>
<td>258</td>
</tr>
<tr>
<td>1</td>
<td>8</td>
<td>0.000</td>
<td>0.074</td>
<td>0.67</td>
<td>0.43</td>
<td>258</td>
</tr>
<tr>
<td>2</td>
<td>5</td>
<td>0.000</td>
<td>0.056</td>
<td>0.62</td>
<td>0.34</td>
<td>258</td>
</tr>
<tr>
<td>2</td>
<td>6</td>
<td>0.014</td>
<td>0.051</td>
<td>0.66</td>
<td>0.41</td>
<td>258</td>
</tr>
<tr>
<td>2</td>
<td>8</td>
<td>0.013</td>
<td>0.049</td>
<td>0.72</td>
<td>0.50</td>
<td>258</td>
</tr>
<tr>
<td>3</td>
<td>5</td>
<td>–0.01</td>
<td>0.054</td>
<td>0.72</td>
<td>0.53</td>
<td>264</td>
</tr>
<tr>
<td>3</td>
<td>6</td>
<td>0.003</td>
<td>0.046</td>
<td>0.75</td>
<td>0.56</td>
<td>264</td>
</tr>
<tr>
<td>3</td>
<td>8</td>
<td>0.002</td>
<td>0.043</td>
<td>0.80</td>
<td>0.65</td>
<td>264</td>
</tr>
</tbody>
</table>
approx. 750–760 mm y⁻¹ because calculated annual rates of ET₀ in 2002–2003 were smaller than those in other years. In 2000–2001 and 2001–2002, only very low deep-percolation rates (<11 mm y⁻¹) were estimated at all plots. In 2002–2003 at plot 8, the annual rate of deep percolation was estimated at approx. 181 mm y⁻¹ due to a high irrigation of 680 mm y⁻¹ that was observed (Tab. 3). In contrast, in 2002–2003, very low deep-percolation rates were estimated at plots 5 and 6 due to low irrigation of 160 mm y⁻¹ applied at the two plots in the previous year (Tab. 3). This indicates that low deep-percolation rates can occur with low rainfall and irrigation not only in the same year, but that it can also continue into the following year.

4 Discussion

4.1 Evaluation of estimated soil water content

The estimation of soil water content resulted in a RMSE varying from 0.043 to 0.074 cm³ cm⁻³ for the three soil layers. Our result can be compared with estimation results by different agro-ecosystem models in which RMSE calculated by comparing estimated with observed volumetric soil water content was within 0.02 and 0.126 cm³ cm⁻³ (Diekkrüger et al., 1995). In a similar study, a model for calculating field-scale water flow was validated, using TDR measurements (Jacques et al., 2002). Here, the RMSE for estimated and measured soil water content was between 0.038 and 0.125 cm³ cm⁻³ (Jacques et al., 2002). In another study, two procedures in a SWB model were used to estimate soil water content (Wegehenkel, 2005). The IA and RMSE between the estimated and TDR-measured soil water content were 0.26–0.78 and 0.028–0.130 cm³ cm⁻³, respectively (Wegehenkel, 2005). The RMSE comparison between this and other results suggests that the variable three-layer model simulates soil water content well, especially for deep soil layers.

In the surface soil layer, the estimated winter soil water content was lower than that measured. One possible reason is that we disregarded the capillary rise of water from the second layer, which may be significant when the topsoil becomes dry owing to a lack of water input (Fig. 3). Overestimating unsaturated hydraulic conductivity can also contribute to the underestimation of soil water content, because very low soil temperature in winter can reduce unsaturated hydraulic conductivity (Giakoumakis and Tsakiris, 1991; Liu et al., 2000). Snowfall in the model is considered directly as equivalent rainfall, and a snowmelt process is not included; as a result, soil evaporation is overestimated under snow-cover conditions. However, in the period from 1998 to 2003, snowfall only accounted for 1.5% of total precipitation, although snowfall frequency was approx. 10.5% of precipitation events. The impact of snowfall on surface water balance should therefore be very small.

4.2 Comparison between the current and previous model

The current SWB model uses the same percolation function as the previous one (Kendy et al., 2003) to redistribute soil

---

**Figure 4:** Seasonal change of lysimeter-measured evapotranspiration (ET₀) and simulated ET₀ at plot 8 in 1998–2003.

**Figure 5:** Comparison between lysimeter-measured evapotranspiration (ET₀) and simulated ET₀ at plot 8.
water content. There are three main improvements in the current model. The first is that soil layers were simplified from the former ten layers to the current three layers, in which the depths of the second and third layers change with the development of rooting depth; the second is that the dual $K_r$-$ET_0$ approach in the current model was used to quantify $ET_a$ instead of the simple pan-evaporation method used in the former; and the third improvement is that the calibration method for $a$ and $K_r$ (the trial-and-error adjustment) in the former model was replaced by the nonlinear least-square Gauss-Newton method in the current model.

Compared to the former model (Kendy et al., 2003), the current model can much better estimate $ET_0$ and $ET_a$. In the former SWB model, $ET_0$ was estimated by measuring class A pan evaporation multiplied by a conversion coefficient of 0.7. However, this conversion coefficient strongly depends on the upwind fetch and on local advection (Rana and Katerji, 2000). In the current model, the FAO-Penman-Monteith method was used to quantify $ET_0$. The main energy-balance component $R_n$ estimated from the FAO-Penman-Monteith method compared well with net radiometer-measured $R_n$ indicated by a RMSE of 2.09 MJ m$^{-2}$ d$^{-1}$, an IA of 0.92, and a CCC of 0.85. Note that the FAO-Penman-Monteith model estimates $R_n$ for grassland, while $R_n$ is measured for cropland of winter wheat and maize.

Compared to the previous study (Kendy et al., 2003), this study improves simulation quality of daily $ET_a$. Between the measured and estimated daily $ET_a$, $R^2$ is approx. 0.74, compared with 0.58 in the previous study (Kendy et al., 2003). The scatter in Fig. 5 may be caused by random errors in both estimated and measured $ET_a$. Incorrect parameterization of crop coefficient and water-stress coefficient in the model can further contribute to the error. The well-known “oasis effect,” mainly reflected by the lysimeter’s metal and concrete frame, can cause higher $ET_a$ rates in the lysimeter than in the surrounding cropland (Zhang et al., 2002).

Under nondeficient-irrigation conditions, the annual $ET_a$ rate estimated by the previous approach (Kendy et al., 2003) was only prox. 680 mm y$^{-1}$, which was much lower than that measured with the large-scale—weighing lysimeter method carried out in the same research station (Liu et al., 2002). These measured $ET_a$ rates ranged between 800 and 900 mm y$^{-1}$. Compared to the pan-evaporation-based $ET_a$ method in the model of Kendy et al. (2003), the dual $K_r$-$ET_0$ approach in the new developed model estimated annual rates of $ET_a$ between 730 and 843 mm at the three plots in 1998–2003. These values are only slightly lower than the lysimeter measurement, which is dubious due to the “oasis effect” discussed above.

The current model estimated soil water content in the first layer better than did the previous model, documented by the fact that the current model produced a RMSE varying between 0.070 and 0.074 cm$^3$ cm$^{-3}$, while the previous one produced 0.101 cm$^3$ cm$^{-3}$ for the first layer (Kendy et al., 2003). We believe that this is due to the improved estimation of $ET_a$ and the better calibration method in the current study.

4.3 Strong and weak points of the current SWB model

In our study, the “tipping bucket” method was applied to redistribute saturated soil water, and an exponential equation was used to estimate unsaturated hydraulic conductivity. The temporal dynamics of the deep-percolation rates estimated by our approach correspond more to natural conditions, compared to some other simple functional SWB models in which soil-water movement is not estimated (Eitzinger et al., 2003; Huang, 2004; Jalota and Arora, 2002). In comparison with similar SWB models (e.g., model SAWAH; Tenberge et al., 1995), our modeling approach needs fewer coefficients in the soil–water flux equations. Compared to approaches using Richard’s equation (Maraux et al., 1998; Serrano, 2004), the solution of the soil water–movement equations in our model needs less computing time because we avoid solving differential equations.

The depths of the middle active soil layer and the lowest passive soil layer in the current model dynamically change according to the development of crop rooting depth. In this case, the accuracy of estimated rooting depth directly influences the precision of estimates of soil moisture in the two layers. In our study, the development of crop rooting depth was estimated based on observation results in one year (Zhang, 1999). However, the real maximum rooting depth in different years might be different due to meteorological conditions and irrigation application, as discussed above. Therefore, the development of rooting depth might vary from year to year. As a result, the interface between the second and third layers might be poorly estimated.

To further simplify the model, we did not estimate snowmelt. Although winter temperatures can dip to $-10^\circ$C at Luancheng station, precipitation during the winter is small. This simplification affects the accuracy of estimated soil water content in the surface soil layer in the winter (Fig. 3), but does not affect deeper soil layers.

5 Conclusions

The current model is useful for calculating the soil water balance in the northern China Plain, where the water table is declining continuously. The soil–water balance estimates (Tab. 3) show that double-cropped winter wheat and summer maize used approx. 750–850 mm y$^{-1}$ soil water under nondeficient-irrigation treatment. Annual rainfall in the study years changed from 347 to 476 mm y$^{-1}$. To maintain this cropping system, at least 250 mm groundwater must be pumped each year. The model simulation suggests that rotating cropland results in the decline of the water table and that winter wheat must be replaced with low water-demanding economic crops.

The current SWB model is particularly suitable to areas with little topographic relief, relatively deep water tables, and insignificant snowmelt such as those in the northern China Plain. The model can provide an appropriate tool for the policymakers in making sound decisions concerning soil–water balance issues in irrigated cropland when crop root depth is accurately estimated.
The estimation accuracy of the SWB model mainly depends on the following conditions: (1) physical parameters of soil profiles to quantify soil water redistribution in a saturated or unsaturated soil layer, (2) crop rooting depth determining the middle active soil layer and the lowest passive soil layer, and (3) the empirical parameter, for quantifying unsaturated hydraulic conductivity.

Acknowledgment

The study was supported by a special foundation by the Chinese Academy of Sciences (Grant No. O7R70020SD).

References


Diekkrüger, B., Sondgerath, D., Kersebaum, K. C., McVoy, C. W. (2007): Saturated hydraulic conductivity of soil profiles to quantify soil water redistribution in a saturated or unsaturated soil layer, (2) crop rooting depth determining the middle active soil layer and the lowest passive soil layer, and (3) the empirical parameter, for quantifying unsaturated hydraulic conductivity.


