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# Frozen soil parameterization in a distributed biosphere hydrological model

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## Abstract

In this study, a frozen soil parameterization has been modified and incorporated into a distributed biosphere hydrological model (WEB-DHM). The WEB-DHM with the frozen scheme was then rigorously evaluated in a small cold area, the Binngou watershed, against the in-situ observations from the WATER (Watershed Allied Telemetry Experimental Research). In the summer 2008, land surface parameters were optimized using the observed surface radiation fluxes and the soil temperature profile at the Dadongshu-Yakou (DY) station in July; and then soil hydraulic parameters were obtained by the calibration of the July soil moisture profile at the DY station and by the calibration of the discharges at the basin outlet in July and August that covers the annual largest flood peak of 2008. The calibrated WEB-DHM with the frozen scheme was then used for a yearlong simulation from 21 November 2007 to 20 November 2008, to check its performance in cold seasons. Results showed that the WEB-DHM with the frozen scheme has given much better performance than the WEB-DHM without the frozen scheme, in the simulations of soil moisture profile at the DY station and the discharges at the basin outlet in the yearlong simulation.

## 1 Introduction

Frozen soil (comprising permafrost and seasonally-frozen soil) process is critically important in the land surface hydrology of cold regions, since the freeze-thaw cycle significantly modulates the soil hydraulic and thermal characteristics that directly affect the water and energy cycles in the soil-vegetation-atmosphere transfer (SVAT) system.

At present, the improved modeling of the frozen soil process in a land surface scheme has been recognized an indispensable task for more reliable estimates of soil moisture and temperature profiles, particularly in the winter of Northern Hemisphere. The Project for Intercomparison of Land surface Parameterization Schemes phase 2(d) (PILPS 2(d)) has shown that the models with a frozen soil parameterization generally

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simulated realistic soil temperature during winter than those without a frozen scheme (Luo et al., 2003). Up to now, many studies have made efforts on improving the frozen soil parameterization in land surface modeling (e.g., Slater et al., 1998; Koren et al., 1999; Smirnova et al., 2000; Li and Koike, 2003; Woo et al., 2004; Niu and Yang, 2006; Zhang et al., 2007; Nicolsky et al., 2007; Luo et al., 2009).

By contrast, the frozen soil process is often inadequately represented or even neglected in most distributed hydrological models for basin-scale simulations, with only very few exceptions (e.g., Cherkauer and Lettenmaier; 1999; Zhang et al., 2000; Tian et al., 2006; Mou et al., 2008). In fact, the representations of the frozen soil process can be indispensable in distributed hydrological modeling for the understanding of the water and energy cycles in some Northern Hemisphere river basins, because many large rivers in Northern Hemisphere (e.g., the Yellow River; see Tang et al., 2008 and the Heihe River; see Gao et al., 2008) originate from high and cold mountain regions. As a result, the hydrological simulations in the upper cold subbasins of these river basins with a spatially-distributed manner become crucial for improving integrated water resources management.

This study aims at improving the performance of a distributed biosphere hydrological model (WEB-DHM; Wang et al., 2009a,b) in the simulations of the frozen soil process, which has been formulated in a previous study (Li and Koike, 2003). This frozen soil scheme will be modified and incorporated into the WEB-DHM for better descriptions of the frozen soil process. By using yearlong continuous observations (soil moisture, soil temperature, and discharge) from the cold region hydrology experiment of WATER (Watershed Allied Telemetry Experimental Research; Li et al., 2008, 2009), the newly-developed WEB-DHM with the revised frozen scheme has been rigorously evaluated in a small cold river basin (Binggou) at the upper area of the Heihe River.

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## 2 Model description

The WEB-DHM (Water and Energy Budget-based Distributed Hydrological Model) was a distributed biosphere hydrological model, which can give consistent descriptions of water, energy and CO<sub>2</sub> fluxes at a basin scale (see Fig. 1). It can efficiently simulate hydrological processes of large-scale river basins while incorporating subgrid topography (see Wang et al., 2009a). In the study, a new version of the WEB-DHM has been developed by modifying and incorporating a frozen soil parameterization (Li and Koike, 2003). First, the surface radiation budgets and the treatments of snow in the WEB-DHM are briefly described in Sects. 2.1 and 2.2. Details about the soil submodel, lateral flow components, and river routing processes were discussed in Wang et al. (2009a); while the formulations of the land surface processes can be found in Sellers et al. (1996a). Second, the frozen soil parameterization in WEB-DHM is presented in detail.

### 2.1 Surface radiation budget

The net amount of radiant energy absorbed by the land surface and available for transfer into other energy forms is defined as the sum of the absorbed solar energy and the absorbed downwelling longwave radiation emitted by the overlying atmosphere minus the longwave radiation emitted by the surface.

$$R_n = R_{sd} - R_{su} + R_{ld} - R_{lu}, \quad (1)$$

where,  $R_n$  is the net radiation;  $R_{sd}$  and  $R_{ld}$  are the downward shortwave and longwave radiation; and  $R_{su}$  and  $R_{lu}$  are upwelling reflected shortwave radiation and longwave radiation emitted by surface, respectively.

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## 2.2 Treatments of snow

First, the ratio of snowfall and rainfall is calculated by following Yamazaki (2001). The ratio of the snow amount to the total precipitation ( $S_{\text{ratio}}$ ) is estimated by

$$S_{\text{ratio}} = 1 - 0.5 \times \exp(-2.2 \times (1.1 - T_W)^{1.3}), \quad (T_W < 1.1^\circ\text{C})$$

$$S_{\text{ratio}} = 0.5 \times \exp(-2.2 \times (T_W - 1.1)^{1.3}), \quad (T_W \geq 1.1^\circ\text{C}), \quad (2)$$

Wet-bulb temperature  $T_W$  can be approximated near  $1^\circ\text{C}$  as

$$T_W = 0.584 \times T_a + 0.875 \times e - 5.32, \quad (3)$$

where,  $T_a$  is air temperature, and  $e$  is water vapor pressure (hPa). As a result, the amount of snowfall is  $S_{\text{ratio}} \times P$  and rainfall is  $(1 - S_{\text{ratio}}) \times P$ , where  $P$  is total precipitation.

Second, the WEB-DHM incorporates a treatment of non-uniform snow cover, which assumes that the snow extent varies linearly with snow water equivalent according to

$$A_s = a_s M_{\text{gs}}, \quad (4)$$

where,  $A_s$  is snow-covered area, and  $0 \leq A_s \leq 1$ ;  $a_s = 13.2 \text{ (m}^{-1}\text{)}$ . The snowpack and its underlying surface soil layer may have a different temperature,  $T_{\text{snow}}$ , than the adjacent bare soil ground within a grid square. The rule is as follows.

$$T_{\text{snow}} = T_g, \quad \text{when } T_g < T_f, \quad (5a)$$

$$T_{\text{snow}} = T_f, \quad \text{when } T_g > T_f. \quad (5a)$$

Where  $T_f$  is the freezing point of water and set to 273.16 K in the model.

Third, once the precipitation has been intercepted by the canopy or ground, the calculations have been made to take account of possible phase changes, using a simple energy balance calculation to calculate a new temperature for the canopy and ground, the intercepted precipitation, and the frozen and melted water produced as a result of phase changes (see Appendix D in Sellers et al., 1996a). At the end of the calculation,

if both liquid and solid water exist, the liquid water is lost and added directly to the surface soil moisture. This prohibits the retention of two different phases of water that would make for complications at the next time step.

Fourth, the effects of snow on surface reflectance and aerodynamic properties have been considered (see Sellers et al., 1996a for details).

## 2.3 Frozen soil parameterization

### 2.3.1 Soil hydraulic properties

Total volumetric water content of the  $j$ th soil layer ( $\theta_j$ ;  $\text{m}^3 \text{m}^{-3}$ ) is defined as

$$\theta_j = \theta_{\text{liq},j} + \theta_{\text{ice},j} \frac{\rho_{\text{ice}}}{\rho_{\text{liq}}} . \quad (6)$$

Where,  $\theta_{\text{liq},j}$  and  $\theta_{\text{ice},j}$  are the liquid water content and the ice content ( $\text{m}^3 \text{m}^{-3}$ ) of the  $j$ th soil layer;  $\rho_{\text{ice}}$  is the density of ice ( $\text{kgm}^{-3}$ );  $\rho_{\text{liq}}$  is the density of liquid water ( $\text{kgm}^{-3}$ ).

For the  $j$ th soil layer, the unfrozen water content ( $\theta_{\text{liq},j}$ ) is assumed as a simple power function of soil temperature ( $T_{\text{soil},j}$ ) (see Nakano et al., 1982; Romanovsky and Osterkamp, 2000; Li and Koike, 2003)

$$\theta_{\text{liq},j} = a(T_f - T_{\text{soil},j})^b , \quad (7)$$

where,  $a$  and  $b$  are two empirical coefficients associated with soil type. Then the changing rate of liquid soil moisture (or ice) to soil temperature can be derived as

$$\frac{\partial \theta_{\text{liq},j}}{\partial T_{\text{soil},j}} = -\frac{\rho_{\text{ice}}}{\rho_{\text{liq}}} \frac{\partial \theta_{\text{ice},j}}{\partial T_{\text{soil},j}} = (-ab)(T_f - T_{\text{soil},j})^{b-1} . \quad (8)$$

The soil hydraulic conductivity  $K_j$  and soil matric potential  $\psi_j$  at the  $j$ th soil layer in the unsaturated zone are described using a modified version of the van Genuchten's

equation (van Genuchten, 1980):

$$K_j = f_{ice,j} K_{sat,j} \left( \frac{\theta_{liq,j} - \theta_r}{\theta_s - \theta_r} \right)^{1/2} \left[ 1 - \left( 1 - \left( \frac{\theta_{liq,j} - \theta_r}{\theta_s - \theta_r} \right)^{-1/m} \right)^m \right]^2, \quad (9)$$

$$\psi_j = \frac{1}{\alpha} \left[ \left( \frac{\theta_{liq,j} - \theta_r}{(\theta_s - \theta_{ice,j}) - \theta_r} \right)^{-1/m} - 1 \right]^{1/n}. \quad (10)$$

Where,  $\theta_s$  is porosity and  $\theta_r$  is residual water content;  $\alpha$ ,  $n$  are empirical parameters in van Genuchten's equation with  $m=1-1/n$ ; the hydraulic conductivity reduction factor for the  $j$ th soil layer ( $f_{ice,j}$ ) is defined as a function of soil temperature at that layer

$$f_{ice,j} = \exp(-10 \times (T_f - T_{soil,j})) \quad \text{and} \quad 0.05 \leq f_{ice,j} \leq 1; \quad (11)$$

and  $K_{sat,j}$  is the saturated hydraulic conductivity at depth  $z$  (defined as the location of the center of the  $j$ th soil layer), which is measured downward in a direction normal to the soil surface (m).  $K_{sat,j}$  is represented using the assumption of an exponential decrease in hydraulic conductivity with increasing soil depth (Beven, 1982; Cabral et al., 1992; Robinson and Sivapalan, 1996)

$$K_{sat,j} = K_{surface} \exp(-fz), \quad (12)$$

where  $K_{surface}$  is the saturated conductivity at the soil surface ( $\text{mmh}^{-1}$ ), and  $f$  is the hydraulic conductivity decay factor.

Similar to  $K_j$ , the groundwater hydraulic conductivity  $K_G$  is also formulated considering the frozen soil effect

$$K_G = f_{ice,G} K_{G0}. \quad (13)$$

Where, the reduction factor  $f_{ice,G}$  ( $=\exp(-10 \times (T_f - T_B))$  and  $0.05 \leq f_{ice,G} \leq 1$ ) is expressed as a function of the temperature of the bottom soil layer ( $T_B$ );  $K_{G0}$  is the groundwater hydraulic conductivity without considering frozen soil.

### 2.3.2 Soil thermal properties

For the surface soil temperature, the force-restore model (Deardorff, 1977) of the heat balance in the soil surface is kept, but the effective heat capacity of surface soil is modified to represent the latent heat of fusion in the surface layer

$$C_g \frac{\partial T_g}{\partial t} = Rn_g - H_g - \lambda E_g - \frac{2\pi C_d}{\tau_d} (T_g - T_d) - \xi_{gs}, \quad (14)$$

where,  $T_g$  and  $T_d$  are the temperature (K) for the soil surface and the deep soil;  $Rn_g$  is the absorbed net radiation by soil surface ( $Wm^{-2}$ );  $H_g$  is the sensible heat flux from soil surface ( $Wm^{-2}$ );  $E_g$  is the bare soil evaporation rate ( $kgm^{-2}s^{-1}$ );  $C_g$  and  $C_d$  are the effective heat capacity ( $Jm^{-2}K^{-1}$ ) for the soil surface and the deep soil;  $\lambda$  is the latent heat of vaporization ( $Jkg^{-1}$ );  $\xi_{gs}$  is the energy transfer due to phase changes in snow on ground ( $Wm^{-2}$ ). The new equation of  $C_g$  is described as

$$C_g = d_s \left( C_s + \rho_{liq} L_f \frac{\partial \theta_{liq,1}}{\partial T_g} \right) + \min[0.05, (M_{gw} + M_{gs})] C_w, \quad (15)$$

where  $d_s$  is the effective depth that feels the diurnal change of temperature (Stull, 1988);  $C_s$  is the volumetric heat capacity of soil ( $Jm^{-3}K^{-1}$ ); the  $\rho_{liq} L_f (\partial \theta_{liq,1} / \partial T_g)$  represents the apparent heat capacity of soil freezing in the surface soil layer;  $M_{gw}$  and  $M_{gs}$  are the soil interception of liquid and snow-ice store, respectively;  $C_w$  is the volumetric heat capacity of water ( $Jm^{-3}K^{-1}$ ).

The depth of seasonal frost penetration is determined by the soil temperature profile, which is solved with Stefan solution (Stefan, 1889). The Stefan solution assumes a linear soil temperature profile in the frozen soil column, but the soil surface temperature usually varies greatly and makes the estimation of soil temperature profile very unstable. As a result, frost/thaw depth is calculated from 5 cm below the surface in the model, where the diurnal change is much smaller and can be assumed as a perfect

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periodic relationship with time. The soil temperature at 5 cm ( $T_{\text{soil},D_1}$ ) is then solved by an analytic solution of the thermal transfer equation (Gao et al., 2000; Li and Cheng, 1995)

$$T_{\text{soil},D_1} = \bar{T}_g - gD_1 + A \exp \left( -D_1 \sqrt{\frac{\omega}{2V_g}} \sin \left( \omega t - D_1 \sqrt{\frac{\omega}{2V_g}} \right) \right), \quad (16)$$

where  $\bar{T}_g$  is the daily mean surface temperature (K);  $g$  is the thermal gradient ( $\text{Km}^{-1}$ ), estimated by the daily averaged temperature difference from the surface to the deep soil;  $A$  is the amplitude of the surface temperature variation (K);  $\omega = 2\pi/l$  and  $l = 86400$  is the period;  $V_g$  is the heat diffusivity ( $\text{m}^2\text{s}^{-1}$ ). Following Li and Koike (2003), the  $T_{\text{soil},D_1}$  can be resolved.

The Frost/thaw depth is simulated following Li and Cheng (1995), and can be expressed in equations of the approximation Stefan solution (The Institute of Geocryology, 1974). After determining the position of freezing front, the soil temperature in the root zone and deep soil zone are estimated by a simple function of frost depth. Details about this part can be found in Li and Koike (2003).

### 3 Datasets for the study area

The cold region hydrology experiment is one of key experiments within the Watershed Allied Telemetry Experimental Research (WATER; Li et al., 2008, 2009). The Binggou watershed (Fig. 2), located at  $100^\circ 12' - 100^\circ 18' \text{ E}$  and  $38^\circ 1' - 38^\circ 4' \text{ N}$ , is one of the three foci experimental areas where cold region hydrology experiments were carried out within the framework of the WATER. The Binggou watershed is a high mountain drainage system with an area of  $30.48 \text{ km}^2$ . It is located in the upper reaches of the Heihe River Basin. The longterm mean annual precipitation is about 686 mm, and snowfall prevails from October to April (Yang et al., 1993). The mean depth of the seasonable snowpack is about 0.5 m, with a maximum of 0.8–1.0 m. The lower limit of permafrost is about 3400 m (Li et al., 2009).

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For the Binggou watershed, the hydrological processes are not only controlled by the hydro-meteorological conditions of the land surface, but also by the underlying frozen soil. In the early spring (April to May), with the air temperature increase, snowmelt occurred from the lower regions to the mountain areas. However, during this period (April to May), the air temperature exceeds 0°C only at noon, but drops a lot at night. Consequently, much of the snowmelt water freezes again at night before its departure from snowpack. Therefore, the snowmelt runoff in the early spring was rather small. From May to June (late spring), the air temperature increases to above 0°C stably, and the snowmelt runoff becomes very large. This is also attributed to the little permeability of the underlying seasonal frozen soil layers which have thawed only in upper soil layers (Kane and Stein, 1983). In summer, snow and seasonal frozen soil layers disappear, and thus rainfall becomes the major source for river discharges. But the permanent frozen soil layers still exist, which prohibit the water infiltration to deeper layers. Heavy rainfall events in summer will usually result in severe flash floods in this watershed, along with landslides and debris flows (Yang et al., 1993).

Digital elevation model (DEM) and land use were provided by the WATER project (Li et al., 2008, 2009). The model simulation adopted a grid size of 250 m, and the subgrid topography was described by a 50 m resolution DEM. The elevation is from about 3400 to 4400 m (Fig. 2). Land is covered by low-density grass, brush, and rock and gravel, and these land uses have been reclassified to 3 SiB2 categories (Agriculture/C3 grassland, Dwarf trees and shrubs, and broadleaf shrubs with bare soil). Soil is the typical frost desert soil.

In the Dadongshu-Yakou (DY) station (see Fig. 2), surface meteorological variables, as well as soil moisture and soil temperature profiles were measured continuously at the 10 min interval from 21 November 2007 to 20 November 2008. The precipitation, relative humidity, air temperature, and wind speed, as well as air pressure, downward longwave and shortwave radiation, obtained from the station were summated into the hourly time series and taken as the forcing data for the whole watershed because the basin is small and distributed data is not available. The surface air temperature inputs

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were modified with a lapse rate of  $6.5\text{Kkm}^{-1}$ , considering the elevation differences between the model grids and the DY station. However, the altitudinal effect on relative humidity was assumed negligible. At the basin outlet, the Binggou stream gauge was newly built in 2008 and discharge data has been obtained from 17 January to 20 November 2008 for model evaluation, with the frequency of several times a day.

The vegetation static parameters including morphological, optical and physiological properties were defined following Sellers et al. (1996b). The dynamic vegetation parameters are Leaf Area Index (LAI), and the Fraction of Photosynthetically Active Radiation (FPAR) absorbed by the green vegetation canopy, and can be obtained from satellite data. Global LAI and FPAR MOD15.BU 1 km data sets (Myneni et al., 1997) were used in this study. These are 8-daily composites of MOD15A2 products and were obtained using the Warehouse Inventory Search Tool (WIST) of NASA.

All simulations were carried out with 250 m spatial resolution and hourly time steps.

## 4 Model evaluations at the Binggou watershed

### 4.1 Model calibration

#### 4.1.1 Parameters

In the summer 2008, the land surface parameters were optimized using the observed surface radiation fluxes and the soil temperature profile at the DY station in July; and then the optimized soil hydraulic parameters were obtained by the calibrations of the July soil moisture profile at the DY station and the discharges at the Binggou gauge in July and August that covers the annual largest flood peak of 2008.

First, the land surface parameters were optimized using the observed surface radiation fluxes and the soil temperature profile at the DY station in July. For the DY station, land is covered by the SiB2 biome 9 (“Agriculture/C3 grassland”). The canopy cover fraction, the height of canopy top, and the height of canopy bottom have been

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designed as 0.9, 0.05 m, 0.005 m, according to the field observations in Li et al. (2009). The soil relectance to visible radiation for the SiB2 biome 9 was optimized as 0.15 using observed upward shortwave radiation in July 2008; while the surface emissivity has been kept as 1.0 following Seller et al. (1996b). The ground roughness length, the root depth, and the top soil depth were optimized as 0.001 m, 0.25 m, and 1.25 m, by matching the simulated and observed soil temperature profile at the DY station. Other time-invariant vegetation parameters were set following Sellers et al. (1996b). Soil hydraulic properties have been kept equal to values derived from FAO (2003) during the calibration of land surface parameters.

Second, the soil hydraulic parameters were optimized by using the measured soil moisture profile at the DY station and the observed discharge at the Binggou gauge. The saturated and residual soil moistures were set as 0.585 and 0.017 according to the long-term soil moisture measurements from 21 November 2007 to 20 November 2008 in the DY station. The van Genuchten parameters  $\alpha$  and  $n$  were optimized as 0.1 and 2.1, respectively, by comparing the simulated and observed soil moisture profile in July 2008 at the DY station. The hydraulic conductivity for groundwater  $K_{G0}$ , the hydraulic conductivity anisotropy ratio  $anik$ , and the hydraulic conductivity decay factor  $f$ , were optimized to obtain good reproduction of the flood event that occurred at the Binggou stream gauge during the summer (July and August) of 2008. The optimization was done using a trial and error method by matching the simulated and observed flood peaks and tails.

The basin-averaged values of the land surface and soil hydraulic parameters used in the Binggou watershed were listed in Table 2.

#### 4.1.2 Calibration results

Figure 3 showed the simulated and observed hourly surface radiation budgets from 1 to 31 July 2008 at DY station with bias error (BIAS) and root mean squared error (RMSE). The observed soil moisture and temperature profiles (see Table 1) were used to initialize the model at the first hour of 1 July 2008; while the initial water table depth

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was assumed as same as the initial depth of the unsaturated zone ( $D_s=1.25\text{m}$ ). After the calibration of a few land surface parameters, the diurnal cycles of surface radiation fluxes can be well represented by the WEB-DHM with acceptable accuracies. The BIAS for simulated upward shortwave radiation ( $R_{su}$ ), upward longwave radiation ( $R_{lu}$ ), and net radiation ( $R_n$ ) at the DY station are  $-2.4\text{Wm}^{-2}$ ,  $-2.5\text{Wm}^{-2}$ , and  $2.0\text{Wm}^{-2}$ , while their RMSE values are  $32.9\text{Wm}^{-2}$ ,  $11.1\text{Wm}^{-2}$ , and  $18.7\text{Wm}^{-2}$ , respectively. It should be mentioned that the direct measurements of  $R_{lu}$  in the station were found erroneous for all periods, comparing to the estimated  $R_{lu}$  from the observed surface soil temperature at 5 cm. The latter one was used as the observed  $R_{lu}$  alternatively in the study. And there are totally 52 h during which the measured soil temperature at the upper layers (0–25 cm) were found problematic during the calibration period, and thus these hours are exempted in the comparisons of  $R_{lu}$  and  $R_n$  as well the soil temperature at surface layer and root zone.

Figure 4 displayed the hourly observed and simulated soil temperature at surface layer (0–5 cm), root zone (5–25 cm), and deep soil (25–125 cm) in July 2008 at the DY station. The soil temperature at surface layer, root zone and deep soil were all well reproduced by the calibrated model with the BIAS of 0.03 K,  $-0.76\text{K}$ , and  $-0.95\text{K}$  and the RMSE of 1.89 K, 1.98 K and 1.21 K, respectively.

Figure 5 illustrated the hourly evolutions of precipitation and the simulated and observed hourly volumetric liquid soil moisture at 5, 10, 20, 40, 80, and 120 cm in July 2008 at the DY station. Reasonable responses of soil moisture at the upper layers (5, 10, and 20 cm) to the rainfall events were reproduced with high accuracies (Fig. 5b–d). The BIAS for simulated soil moisture at 5, 10, and 20 cm were 0.001, 0.000, and 0.016; while their RMSE values were 0.030, 0.031 and 0.037, respectively. The soil moisture at 40, 80 and 120 cm were also well simulated by the WEB-DHM with the frozen scheme (see Fig. 5e–g). The BIAS values for the simulated soil moisture at 40, 80 and 120 cm were  $-0.036$ ,  $-0.024$ , and 0.005; while their RMSE values were 0.052, 0.067 and 0.007, respectively.

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Figure 6 plotted the calibrated hourly hydrograph from July to August 2008 at the Binggou gauge. It was shown that after calibration, the model reproduced both the peak flow and base flows very well. It should be mentioned that the discharges were measured discontinuously and irregularly, and thereby the evaluation criterions (e.g., the Nash-Sutcliffe model efficiency coefficient (Nash and Sutcliffe, 1970) and the bias error) were not presented here.

## 4.2 Model validation

The calibrated WEB-DHM with the frozen scheme was then used for a yearlong simulation from 21 November 2007 to 20 November 2008, to check its performance in cold seasons.

### 4.2.1 Soil temperature at the DY station from 21 November 2007 to 20 November 2008

Figure 7 gave the hourly observed and simulated soil temperature at surface layer, root zone, and deep soil from 21 November 2007 to 20 November 2008 at the DY station, by using the WEB-DHM with the frozen scheme. In general, the WEB-DHM with the frozen scheme well reproduced the yearlong soil temperature at different soil layers, with the BIAS of 1.79 K, 1.77 K, and 2.40 K and the RMSE of 3.53 K, 3.45 K, and 3.05 K for the surface layer, root zone, and deep soil, respectively. The much larger diurnal changes of soil temperature than the observed ones (measured with heat flux transducer) from January to March 2008, simulated by the WEB-DHM with the frozen scheme, was possibly caused by the underestimation of snow accumulation on the ground.

### 4.2.2 Soil moisture at the DY station from 21 November 2007 to 20 November 2008

Figure 8 illustrated the hourly precipitation, as well as the simulated and observed hourly volumetric liquid soil moisture at 5, 10, 20, 40, 80, and 120 cm from 21 November

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2007 to 20 November 2008 at the DY station. Results showed that the WEB-DHM with the frozen scheme generally gave more realistic yearlong soil moisture profile than those simulated by the WEB-DHM without the frozen scheme. For the soil moisture at 5, 10, 20, 40, and 80 cm, the results by the WEB-DHM with the frozen scheme obtained the RMSE of 0.092, 0.077, 0.057, 0.086, and 0.108, respectively; while the ones by the WEB-DHM without the frozen scheme got the RMSE of 0.104, 0.148, 0.140, 0.161, and 0.185, respectively. The overestimation of soil moisture at 120 cm, from May to November 2008 by the WEB-DHM with the frozen scheme was possibly attributed to the simulated deeper thawing front than the actual one.

**4.2.3 Discharges at the Binggou gauge from 17 January to 20 November 2008**

By using the measured streamflows from 17 January to 20 November 2008, the simulated longterm (including cold seasons) hourly discharges at Binggou gauge were further evaluated.

Figure 9 displayed the hourly hydrographs simulated by the WEB-DHM with and without the frozen scheme at the Binggou gauge from 17 January to 20 November 2008. Results showed that the streamflows simulated by the WEB-DHM with the frozen scheme agreed fairly well with the observations; while those calculated by the WEB-DHM without the frozen scheme had large difference from the observations, with the overestimation of baseflows (January–April), the underestimation of snowmelt flows (April–May), and the overestimation of peak flows (July–August). The improved simulation results by the WEB-DHM with the frozen scheme can be attributed to the following reasons:

1. The consideration of reduction factor of the groundwater hydraulic conductivity ( $f_{ice,G}$ ) largely improved the model's performance from January to April (also see Fig. 10). Figure 10a demonstrated the logarithmic hourly hydrographs simulated by using the WEB-DHM with the frozen scheme at the Binggou gauge from 17

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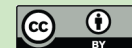
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January to 20 November 2008, which further confirmed the good performance of the new system in simulating base flows. Without the treatments of the frozen soil effect on the groundwater hydraulic conductivity, poorer baseflows have been obtained from January to April 2008 (see Fig. 10b).

2. In the spring season (April–May), more snowmelt runoff calculated by the WEB-DHM with the frozen scheme was due to the treatments of the hydraulic conductivity reduction for each soil layer ( $f_{ice,j}$ ) as a function of soil temperature, which has resulted in less infiltration during the snow melting.
3. The poorer flood peaks obtained by the WEB-DHM without the frozen scheme, were possibly caused by the overestimation of soil moisture at upper soil layers during the flood season (also see Fig. 8b and c).

## 5 Concluding remarks

In this study, the distributed biosphere hydrological model WEB-DHM was improved by incorporating a frozen soil parameterization. The WEB-DHM with the frozen scheme was then applied to the Binggou watershed for evaluation using the in-situ observations from WATER.

After calibrating land surface parameters and soil hydraulic parameters against summer data, the WEB-DHM with the frozen scheme was used for a yearlong validation from 21 November 2007 to 20 November 2008, to check the model's applicability to cold seasons. Results showed that the WEB-DHM with the frozen scheme has given much better performance than the WEB-DHM without the frozen scheme, in the simulations of soil moisture profile at the DY station and the discharges (base and peak flows as well as snowmelt runoff) at the basin outlet in the yearlong simulation.

Different from Li and Koike (2003) that formulated frozen soil process in a 1-D land surface model (SiB2), this study has modified and incorporated the frozen soil scheme into a distributed biosphere hydrological model (WEB-DHM). The newly-developed



WEB-DHM with the frozen scheme has made it possible to simulate the basin-scale cold-region land surface hydrological processes in a spatially-distributed manner while considering the topographically-driven lateral flows. It can be used as a model operator in the catchment-scale land surface hydrological data assimilation system in cold regions, to improve modelling of soil moisture and the surface energy budget as well as streamflows.

*Acknowledgements.* This study was funded by grants from Japan Aerospace Exploration Agency. Parts of this work were also support by grants from the Ministry of Education, Culture, Sports, Science and Technology of Japan. The data used in the paper are provided by the Chinese Academy of Sciences Action Plan for West Development Program (grant number: KZCX2-XB2-09) and Chinese State Key Basic Research Project (grant number: 2007CB714400). The fifth author (H. Li) was supported by National Natural Science Foundation of China (grant number: 40671040).

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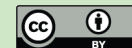
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**Table 1.** Initial conditions for the point-scale simulations at the DY station from 1 to 31 July 2008.

Symbol	Parameters	Unit	Value	Source
$T_c$	Canopy temperature	K	277.43	Li et al. (2009)
$T_g$	Soil surface temperature	K	277.43	Li et al. (2009)
$T_d$	Deep soil temperature	K	276.95	Li et al. (2009)
$\theta_{L,1}$	Volumetric liquid soil moisture at 0.00~0.05m		0.248	Li et al. (2009)
$\theta_{L,2}$	Volumetric liquid soil moisture at 0.05~0.15m		0.226	Li et al. (2009)
$\theta_{L,3}$	Volumetric liquid soil moisture at 0.15~0.25m		0.251	Li et al. (2009)
$\theta_{L,4}$	Volumetric liquid soil moisture at 0.25~0.35m		0.251	Linear Interpolation
$\theta_{L,5}$	Volumetric liquid soil moisture at 0.35~0.45m		0.251	Li et al. (2009)
$\theta_{L,6}$	Volumetric liquid soil moisture at 0.45~0.55m		0.215	Linear Interpolation
$\theta_{L,7}$	Volumetric liquid soil moisture at 0.55~0.65m		0.179	Linear Interpolation
$\theta_{L,8}$	Volumetric liquid soil moisture at 0.65~0.75m		0.142	Linear Interpolation
$\theta_{L,9}$	Volumetric liquid soil moisture at 0.75~0.85m		0.106	Li et al. (2009)
$\theta_{L,10}$	Volumetric liquid soil moisture at 0.85~0.95m		0.093	Linear Interpolation
$\theta_{L,11}$	Volumetric liquid soil moisture at 0.95~1.05m		0.079	Linear Interpolation
$\theta_{L,12}$	Volumetric liquid soil moisture at 1.05~1.15m		0.066	Linear Interpolation
$\theta_{L,13}$	Volumetric liquid soil moisture at 1.15~1.25m		0.052	Li et al. (2009)

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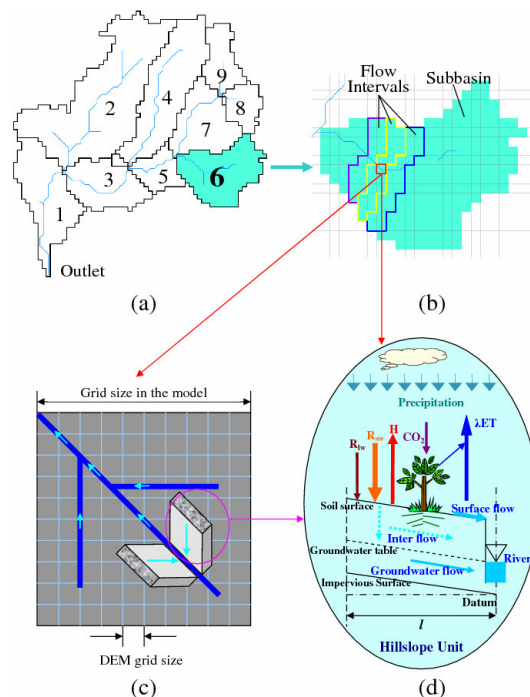


**Table 2.** Basin-averaged values of the parameters used in the Binggou watershed.

Symbol	Parameters	Unit	Value	Source
Land surface parameters				
$z_2$	Height of canopy top	m	0.05	Li et al. (2009)
$z_1$	Height of canopy bottom	m	0.005	Li et al. (2009)
$V$	Canopy cover fraction		0.3	Li et al. (2009)
$\alpha_{s,v}$	Soil reflectance to visible radiation		0.12	Optimization
$z_s$	Ground roughness length	m	0.001	Optimization
$D_r$	Root depth ( $D_1+D_2$ )	m	0.25	Optimization
$D_s$	Top soil depth	m	1.25	Optimization
Soil hydraulic parameters				
$\theta_s$	Porosity		0.585	Li et al. (2009)
$\theta_r$	Residual water content		0.017	Li et al. (2009)
$K_{\text{surface}}$	Saturated hydraulic conductivity for soil surface	mm/h	22	FAO (2003)
$f$	Hydraulic conductivity decay factor		1.84	Optimization
$\alpha$	van Genuchten parameter		0.1	Optimization
$n$	van Genuchten parameter		2.1	Optimization
$anik$	Hydraulic conductivity anisotropy ratio		11.2	Optimization
$K_{G0}$	Hydraulic conductivity for groundwater	mm/h	1.0	Optimization

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**Fig. 1.** Overall structure of WEB-DHM: **(a)** division from basin to sub-basins, **(b)** subdivision from sub-basin to flow intervals comprising several model grids, **(c)** discretization from a model grid to a number of geometrically symmetrical hillslopes, and **(d)** description of the water moisture transfer from atmosphere to river. Where  $R_{sw}$  and  $R_{lw}$  are downward solar radiation and longwave radiation, respectively,  $H$  is the sensible heat flux, and  $\lambda$  is the latent heat of vaporization. Here, the land surface submodel is used to describe the transfer of the turbulent fluxes (energy, water, and  $\text{CO}_2$ ) between atmosphere and ground surface for each model grid; while the hydrological submodel simulates both surface and subsurface runoff using grid-hillslope discretization, and then simulates flow routing in the river network.

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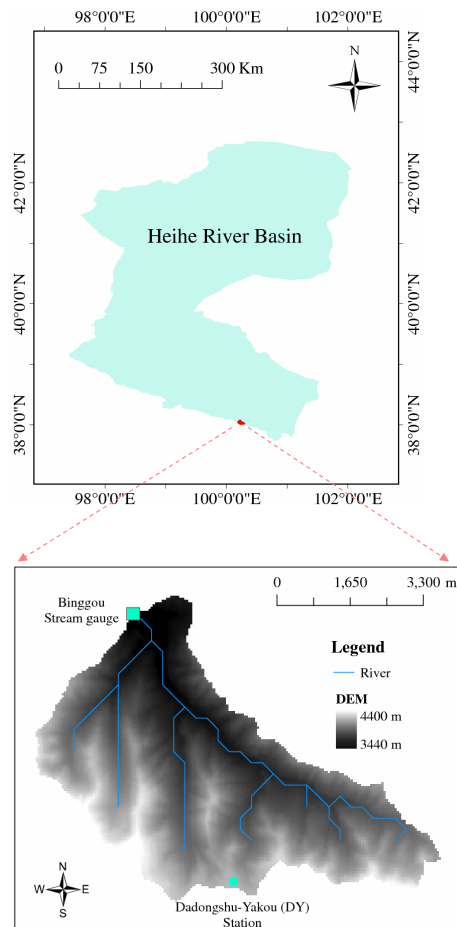
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**Fig. 2.** The Binggou watershed.

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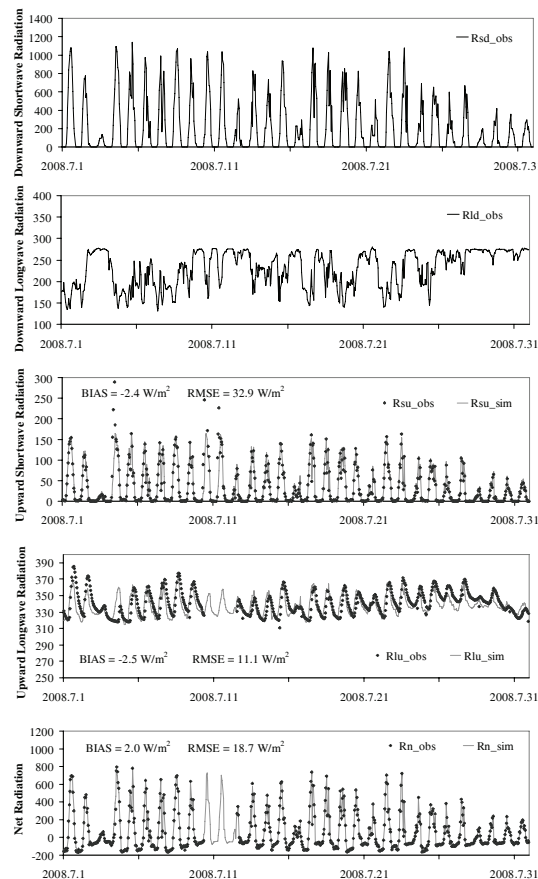
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**Fig. 3.** Simulated and observed hourly upward shortwave and longwave radiation, as well as net radiation (unit:  $W/m^2$ ) in July 2008 at the DY station, by using the WEB-DHM with the frozen scheme.

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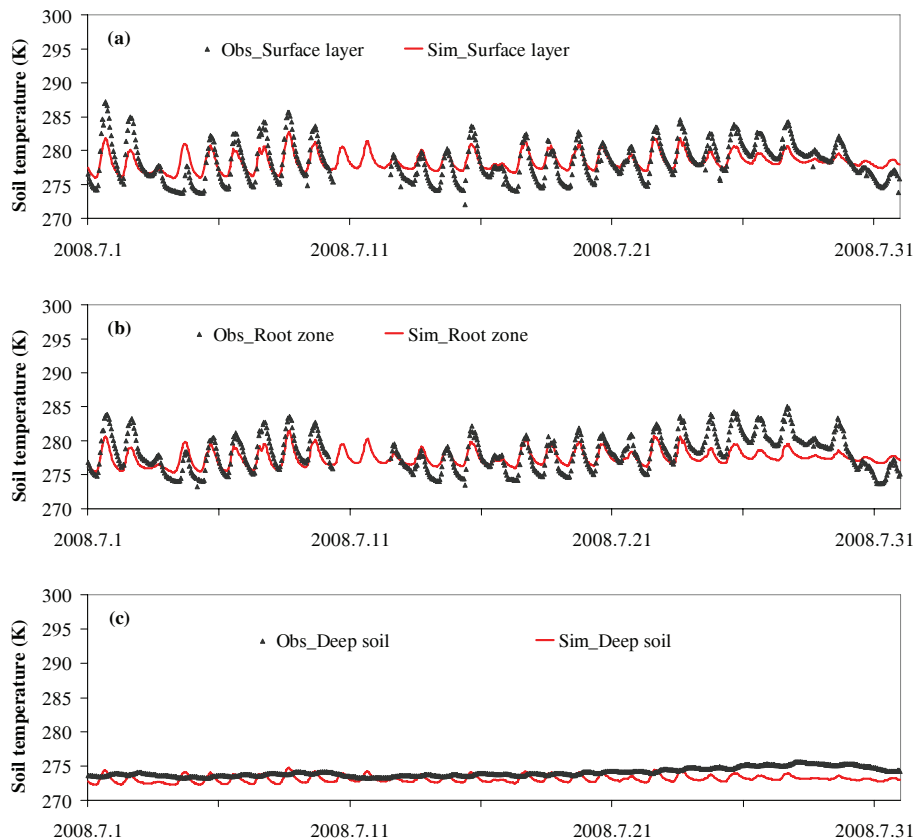
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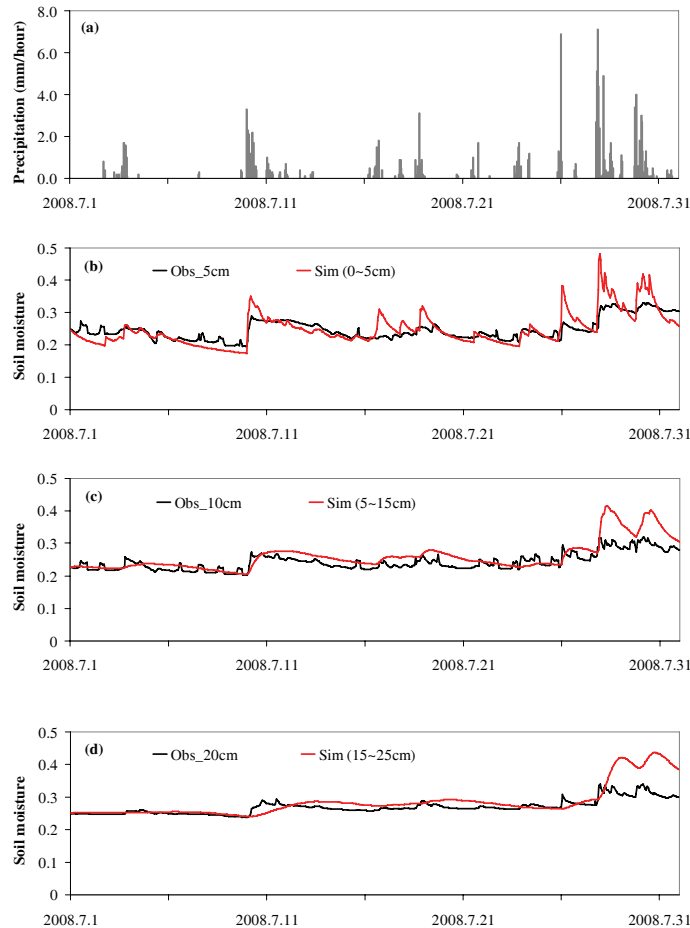
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**Fig. 4.** Hourly observed and simulated soil temperature at surface layer (0–5 cm), root zone (5–25 cm), and deep soil (25–125 cm) in July 2008 at the DY station, by using the WEB-DHM with the frozen scheme.

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**Fig. 5a–d.** Hourly precipitation **(a)**, and the simulated and observed hourly volumetric liquid soil moisture at 5, 10, 20, 40, 80, and 120 cm **(b–g)** in July 2008 at the DY station, by using the WEB-DHM with the frozen scheme.

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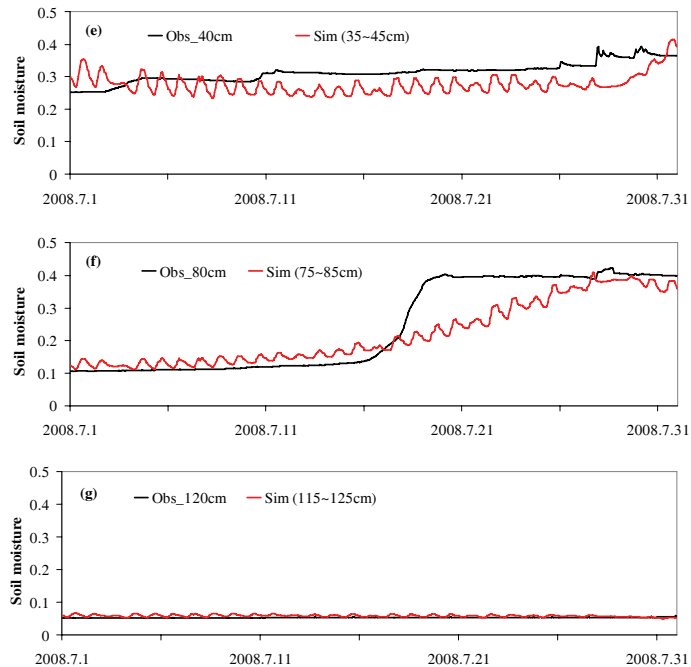
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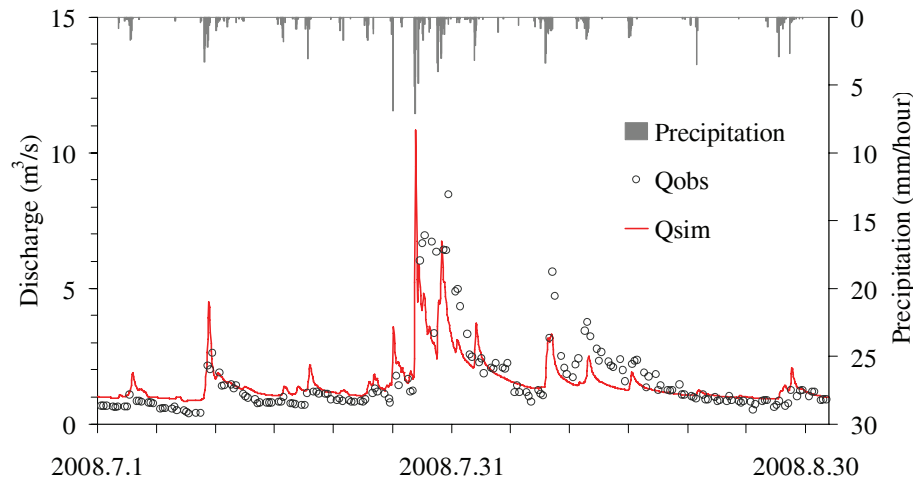
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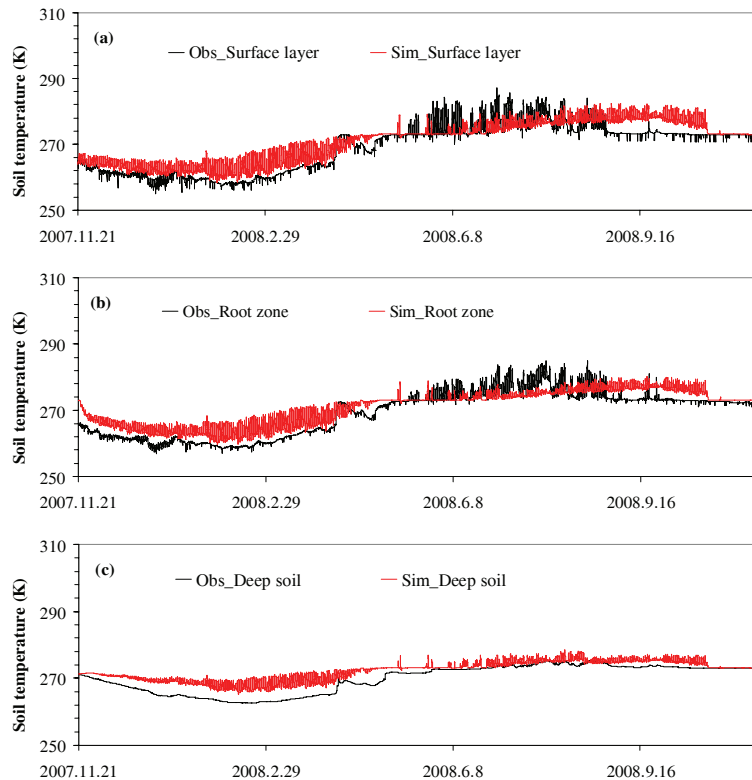


**Fig. 6.** Simulated and observed hourly streamflows at the Binggou gauge from 1 July to 31 August 2008 by using the WEB-DHM with the frozen scheme.

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**Fig. 7.** Hourly observed and simulated soil temperature at surface layer (0–5 cm), root zone (5–25 cm), and deep soil (25–125 cm) from 21 November 2007 to 20 November 2008 at the DY station, by using the WEB-DHM with the frozen scheme.

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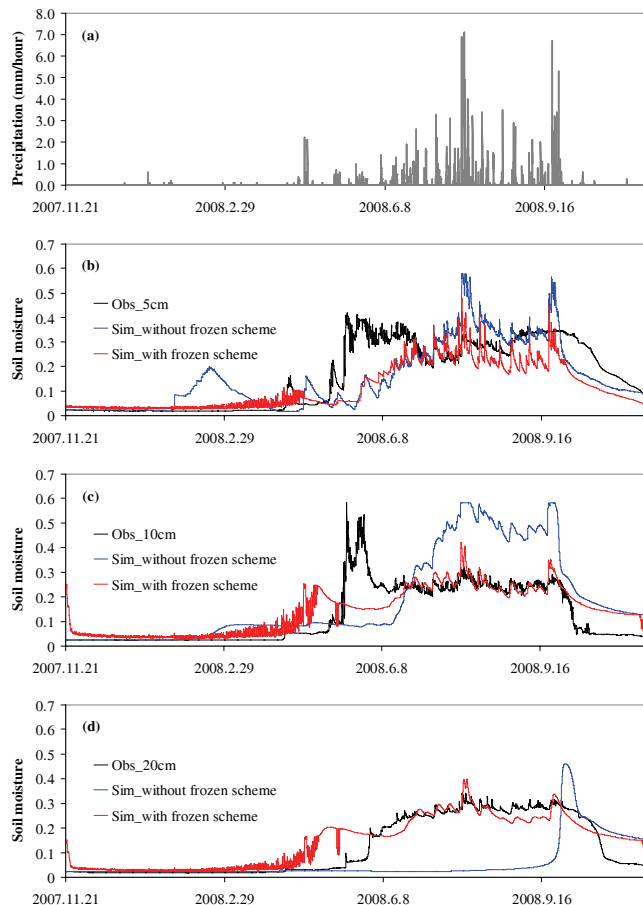
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**Fig. 8a–d.** Hourly precipitation (a), and the simulated and observed hourly volumetric liquid soil moisture at 5, 10, 20, 40, 80, and 120 cm (b–g) from 21 November 2007 to 20 November 2008 at the DY station, by using the WEB-DHM with and without the frozen scheme.

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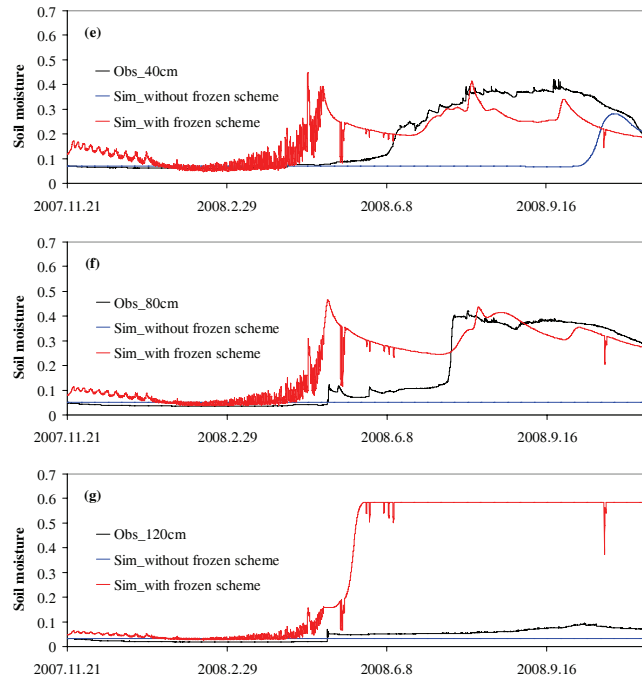


Fig. 8e–g. Continued.

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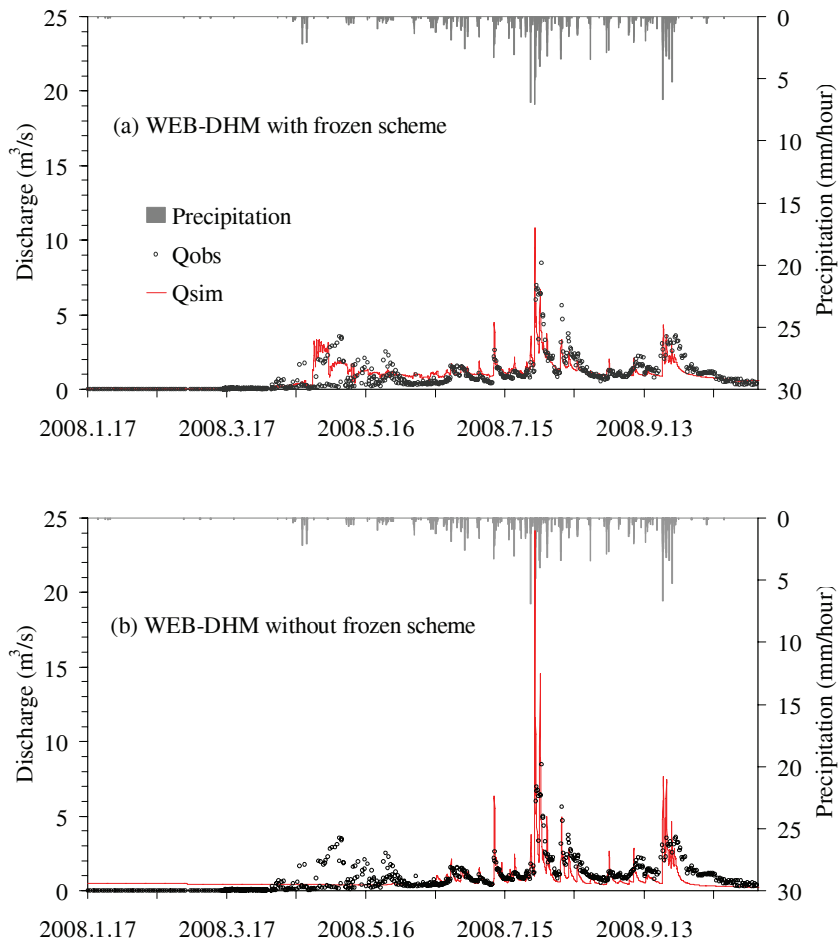
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**Fig. 9.** Linear hourly hydrographs simulated by the WEB-DHM with and without the frozen scheme at the Binggou gauge from 17 January to 20 November 2008.

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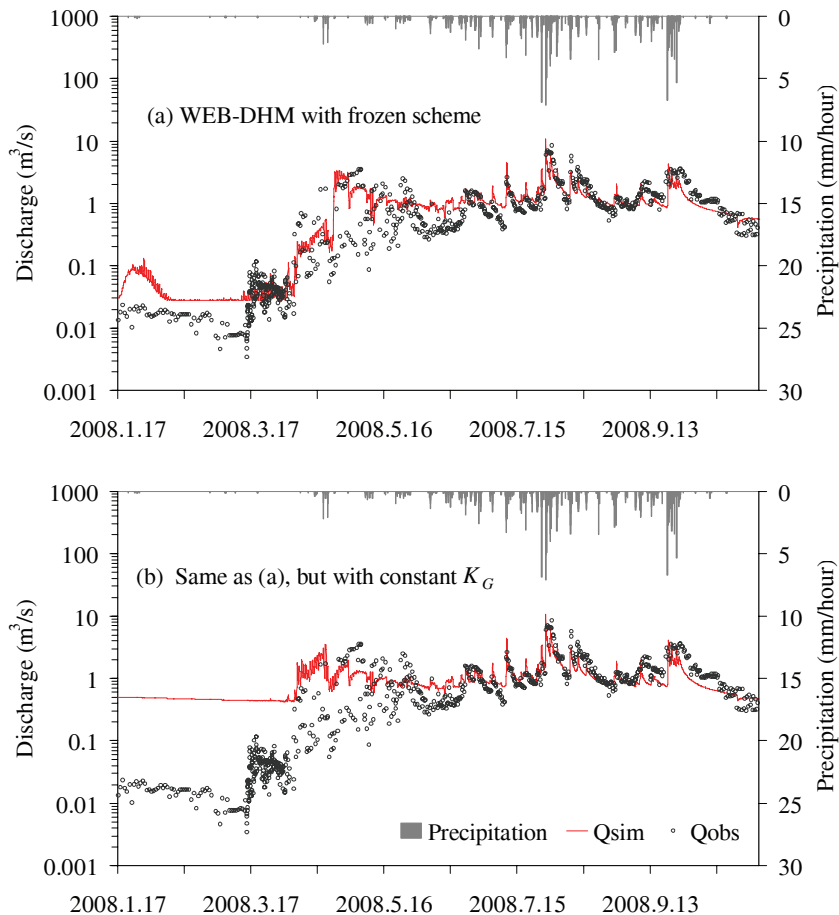
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**Fig. 10.** Logarithmic hourly hydrographs simulated by using the WEB-DHM with the frozen scheme (a), and the model with constant groundwater hydraulic conductivity at the Binggou gauge from 17 January to 20 November 2008.

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