

A distributed water-heat coupled model for mountainous watershed of an inland river basin of Northwest China (I) model structure and equations

Ren-sheng Chen · Shi-hua Lu · Er-si Kang · Xi-bin Ji ·
Zhihui Zhang · Yong Yang · Wenwu Qing

Received: 1 November 2006 / Accepted: 12 March 2007 / Published online: 20 April 2007
© Springer-Verlag 2007

Abstract It is absolutely necessary to quantify the hydrological processes in earth surface by numerical models in the cold regions where although most Chinese large rivers acquire their headstreams, due to global warming, its glacier, permafrost and snow cover have degraded seriously in the recent 50 years. Especially in an arid inland river basin, where the main water resources come from mountainous watershed, it becomes an urgent case. However, frozen ground's impact to water cycle is little considered in the distributed hydrological models for a watershed. Took Heihe mountainous watershed with an area of 10,009 km², as an example, the authors designed a distributed heat-water coupled (DWHC) model by referring to SHAW and COUP. The DWHC model includes meteorological variable interception model, vegetation interception model, snow and glacier melting model, soil water-heat coupled model, evapotranspiration model, runoff generation model, infiltration model and flow concentration model. With 1 km DTM grids in daily scale, the DWHC model describes the basic hydrological processes in the research watershed, with 3–5 soil layers for each of the 18 soil types, 9 vegetation types and 11 landuse types, according to the field measurements, remote sensing data and some previous research results. The model can compute the continuous equation of heat and water flow in the soil and can estimate them continuously, by numerical methods or by some empirical formula, which depends on freezing soil status. However, the model still has some

conceptual parameters, and need to be improved in the future. This paper describes only the model structure and basic equations, whereas in the next papers, the model calibration results using the data measured at meteorological stations, together with Mesoscale Model version 5 (MM5) outputs, will be further introduced.

Keywords Frozen soil · Soil temperature · Soil water content · Soil water tension

Nomenclature

a	adjustable parameter for soil evaporation
a_1	empirical constants (clay: 0.13; sand type soil: 0.1)
a_2	empirical constant (clay: -0.129; sand type soil: 0.058)
a_3	empirical constant (0.6245 for both clay and sand type soil)
b	adjustable parameter for vegetation transpiration
b_1	empirical constants (clay: 0.00144; sand type soil: 0.00158)
b_2	empirical constant (clay: 1.32; sand type soil: 1.336)
b_3	empirical constant (clay: 0.0036; sand type soil: 0.00375)
b_4	empirical constant (clay: 0.8743; sand type soil: 0.9118)
C	soil heat capacity (J °C ⁻¹ cm ⁻³)
C_i	specific heat of ice (J g ⁻¹ °C ⁻¹)
C_l	specific heat of water (J g ⁻¹ °C ⁻¹)
C_s	specific heat of soil grains (J g ⁻¹ °C ⁻¹)
d_1	empirical constant (default 0.5)
d_2	empirical constants (default 0.1)
d_3	empirical constants (default 10)

R. Chen (✉) · S. Lu · E. Kang · X. Ji · Z. Zhang ·
Y. Yang · W. Qing
Cold and Arid Regions Environmental and Engineering
Research Institute, Chinese Academy of Sciences,
Lanzhou 730000, China
e-mail: crs2008@lzb.ac.cn

E	evapotranspiration in a DTM grid (mm)	R_0	surface runoff (mm d ⁻¹)
E_0	potential evaporation (mm)	R_{glacier}	glacier meltwater (mm d ⁻¹)
E_c	evaporation of canopy interception (mm)	R_I	subsurface flow of soil layer I (mm d ⁻¹)
E_s	soil evaporation (mm)	R_{snow}	snow meltwater (mm d ⁻¹)
E_v	vegetation transpiration (mm)	T	air temperature
f_{lat}	ratio of latent heat of ice to the total heat content of the soil, $Q_{\text{total},tf}$, at the temperature T_f	t	flow time between two adjacent DTM grids
g	gravity constant (9.8 ms ⁻²)	T_0	soil surface temperature
H	sensitive heat content (J cm ⁻² d ⁻¹)	T_f	a threshold temperature value below which the soil is assumed to be completely frozen except of a residual unfrozen amount (−8°C)
h_1	empirical constants (default 0.06)	T_s	soil temperature (°C)
h_2	empirical constants (default 0.01)	$T_{s,\text{change}}$	increase or decrease of soil temperature (°C)
h_3	empirical constant (default 2.0)	V_{cov}	vegetation coverage
I–III	soil layer number	V_p	precipitation that intercepted by vegetation canopy (mmH ₂ O)
k'_0	corrected k_0 considering frozen effects (cm d ⁻¹)	$V_{p,\text{total}}$	canopy storage amount (mmH ₂ O)
k_0	saturated hydraulic conductivity of unfrozen soil (cm d ⁻¹)	V_{p0}	vegetation interception capacity (mmH ₂ O)
k_h	thermal conductivity for unfrozen soil (J s ⁻¹ m ⁻¹ °C ⁻¹)	$V_{p\text{max}}$	saturated interception capacity (mmH ₂ O)
$k_{h,i}$	thermal conductivity for fully frozen soil (J s ⁻¹ m ⁻¹ °C ⁻¹)	V_{sto}	canopy storage amount before one precipitation (mmH ₂ O)
$k_{h,s}$	thermal conductivity for sub-frozen soil (J s ⁻¹ m ⁻¹ °C ⁻¹)	w	soil total mass of water (g)
k_{hm}	thermal conductivity for unfrozen mineral soil (J s ⁻¹ m ⁻¹ °C ⁻¹)	w_{ice}	mass of water available for freezing (g)
$k_{hm,i}$	thermal conductivity of fully frozen mineral soil (J s ⁻¹ m ⁻¹ °C ⁻¹)	z	soil layer depth (cm)
k_{ho}	thermal conductivity for unfrozen humus (J s ⁻¹ m ⁻¹ °C ⁻¹)	λ	pore size distribution index
$k_{ho,i}$	thermal conductivity of a fully frozen organic soil (J s ⁻¹ m ⁻¹ °C ⁻¹)	β	slope of a DTM grid (%)
k_{mat}	saturated matrix conductivity (cm d ⁻¹)	θ	soil volume water content (%)
k_w	unsaturated hydraulic conductivity (cm d ⁻¹)	θ_c	field capacity (%)
l	flow length in a DTM grid (m)	θ_l	soil volume liquid water content (%)
LAI	leaf area index	θ_m	soil volume water content at ψ_{mat} , $\theta_m = \theta_s - 4\%$ (%)
LAI _{max}	maximal LAI in a year	θ_i	soil volume solid water content (%)
L_f	latent heat of freezing (334 × 10 ³ J kg ⁻¹)	$\theta_{i,\text{change}}$	mass change of the solid water content (g)
m_1	adjustable parameter for flow time	θ_r	residual water content (%)
m_2	adjustable parameter for flow time	θ_s	saturated volume water content (porosity,%)
n	parameter accounting for pore correlation and flow path tortuosity (default 1)	θ_{wilt}	water content at wilting point (%), defined as a tension of 15,000 cm water, i.e. ψ_{wilt}
P	daily precipitation (mmH ₂ O)	θ_x	threshold volume water content (%) at the threshold tension, ψ_x
P_{ground}	precipitation falling to the ground after the interception process (mmH ₂ O)	ρ_d	soil dry bulk density (g cm ⁻³)
Q	thermal quality of the soil layer	ρ_G	soil grains density (g cm ⁻³)
q	infiltration rate (cm d ⁻¹)	ρ_i	ice density (g cm ⁻³)
Q_{total}	heat content of sub-frozen or unfrozen soil (J cm ⁻¹)	ρ_l	liquid water density (g cm ⁻³)
ΔQ_{total}	change of the total heat content of the soil layer (J cm ⁻¹)	ρ_s	soil density (g cm ⁻³)
$Q_{\text{total},tf}$	total heat content of soil at T_f (J cm ⁻¹)	ψ	soil water tension (cmH ₂ O)
r	freezing-point depression	ψ'	corrected water tension for sub-frozen or fully frozen soil layer (cmH ₂ O)
		ψ_a	air entry tension (cmH ₂ O)
		ψ_{mat}	matrix tension (cmH ₂ O)
		ψ_{wilt}	threshold tension (15,000 cmH ₂ O) at wilting point
		ψ_x	threshold tension (cmH ₂ O)

Introduction

Cold regions in China, with an area of 417.4×10^4 km², covering 43.5% of the country's land area (Chen et al. 2006a), are relatively abundant in solid and liquid precipitation (Yang et al. 2000b), and are regarded as the primary headsprings of water resources for Chinese arid and semiarid regions. However, under global warming today, the glacier shrinks seriously and the permafrost degenerates universally in the cold regions in China, which therefore alters their ecology and microclimate environment, resulting in the changes of watershed's runoff generation processes and its total runoff amount in the cold regions in China. Glacier hydrology, snow hydrology and permafrost hydrology make up the hydrology in the cold regions. Comparatively, the hydrological processes of the seasonal frozen soil or permafrost can be widely seen in the cold regions in China, and therefore it deserves more attention. During the runoff generation processes in a watershed of the cold regions, the frozen front soil with variational depths, containing no air-filled pore, is relatively impermeable. Compared with unfrozen soil, frozen or thawing latent heat component should be considered for both seasonal frozen soil and permafrost, due to heat transfer processes in frozen soil. At the same time, the frozen soil's thermal conductivity, specific heat and total heat content, functions of the solid water content, are different from those when it is unfrozen or melted, for the same soil sample/layer. Moreover, compared with the unfrozen soil, the existence of ice in frozen soil changes its soil structure, such as effective porosity, effective field capacity, so do its effective hydraulic conductivity and its soil water potential–water content relationship/characteristic curve of soil moisture, which therefore alter the soil water transfer processes. In a watershed of the cold regions, the land surface area is the mutual interface of the atmosphere and pedosphere, thus the soil water–heat coupled process should be involved in the infiltration, runoff generation, and evapotranspiration processes. Therefore, to know about the water cycle in the cold regions in China, the hydrological processes in frozen soil should be observed and simulated.

Although the hydrology of frozen soils for the most part has been studied experimentally, there are only a few examples of numerical simulation models that describe water infiltration into frozen soils in detail. Most existing soil water transfer models do not include frost effects (Bayard and Stähli 2005). One of the first models coupling heat and water fluxes for a layered soil profile was developed by Harlan (1973). Recently, the models SOIL (Jansson and Halldin 1979), which was further developed and renamed COUP (Jansson and Karlberg 2001), and

SHAW (Flerchinger and Saxton 1989) extended Harlan's concept to detailed soil–vegetation–atmosphere transfer (SVAT) schemes including a process description of soil freezing and thawing, as well as formation and ablation of the snow cover. Since these models treat only one-dimensional water flow, they have been predominately tested ‘‘at the plot scale’’ (Bayard and Stähli 2005). However, some applications of these models were run in a quasi 2D mode, representing hillslope runoff with a sequence of coupled profiles (Stähli et al. 2001). Larger-scale models that include soil frost processes are mostly conceptual (Bayard and Stähli 2005). Ippisch (2001) implemented a soil frost routine in a 3D water and heat flow model, but the coupling between the soil and the snow cover was not integrated.

In an inland river basin of Arid Regions of Northwest China (ARNC), the runoff out from mountainous watershed is the main water resources (Kang et al. 1999). Therefore, it is very important to simulate and to predict the flow discharge in mountainous watershed for its sustainable development in the mid- and downstream regions, where the precipitation is very rare. In the mountainous watershed of most inland river basin in ARNC, the total glacier area is small although glacier number is much, thus the glacier runoff component is little. Similarly, the precipitation in mountainous area is usually in solid form, but seasonal snow cover can be found in most parts of the area. Hence, frost ground impact to water cycle is very important in these regions. Therefore, a simple distributed water–heat coupled model including soil freezing and thawing processes, with some conceptual parameters due to lack of detailed meteorological, hydrological and soil data in ARNC, should be built up. The model should (1) input little meteorological variable and be relatively simple; (2) include primary hydrological processes in cold regions, especially the hydrological processes in the frozen soil; (3) meet the measured runoff series to some extent at the present time. Based on the initial distributed hydrological model in ARNC (Chen et al. 2003), with some methods referred to SWAT (Arnold et al. 1998), SHAW (Flerchinger and Saxton 1989) and COUP (Jansson and Karlberg 2001), a distributed water–heat coupled (DWHC) model for the mountainous watershed of an inland river basin in ARNC was set up. Due to lacking sufficient hydrometeorological data, detailed soil information and field and indoor experiments, the DWHC is very coarse and need to be improved endlessly.

The DWHC model was applied in the mountainous watershed of Heihe mainstream basin. Heihe river, originated from the Qilian Mountains, running through the Hexi Corridor and scattering in the deserts, is one of the largest inland rivers in ARNC, with a drainage area of

130,000 km². The mountainous watershed of Heihe river basin, ranging from 1,674 m to 4,823 m in altitude, has a catchment area of 10,009 km², where the glacier area is 59 km². The averaged yearly runoff out from this mountainous watershed is about 15.5×10^8 m³ (1,944 ~ 2,004, Yingluoxia station, 38° 48'N, 100°11'E, 1,674 m), among which the yearly glacier meltwater is 0.546×10^8 m³ (Kang et al. 1999), accounting for only 3.5% of the total runoff amount. The climate of the mountainous watershed has been described in Chen et al. (2006b) in detail.

Model structure

The DWHC model is composed of eight sub-models (Fig. 1). It is designed according to the water transfer processes, in which snow/glacier melting processes and canopy interception processes are considered. The soil water–heat coupled model deriving from COUP (Jansson and Karlberg 2001) is linked with the evapotranspiration, infiltration and runoff production processes. The model is driven by two types of meteorological data, one comes from the surface meteorological stations, another is the outputs from MM5 (Gao et al. 2004). Contents on how to use these two types of meteorological data to drive the DWHC model would be described in the next papers: (II) using the measured data at the meteorological & hydrological stations; (III) using the outputs from Mesoscale Model version 5.

Spatial and temporal resolution and input variables

Spatial and temporal resolution

Based on Digital Terrain Model (DTM) in Alberts Equal Area Conic projection (Yang et al. 2000a) deriving from topographic map in 1:100,000 scale, and, with the

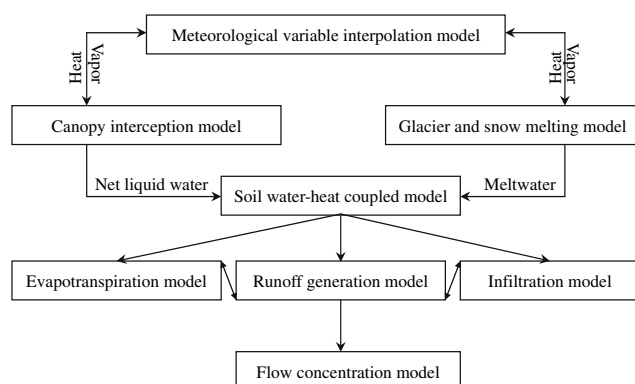


Fig. 1 Model structure and flowsheet

consideration of rare meteorological data & soil research information in the large Heihe mountainous watershed area (10,009 km²), the research watershed is divided into 1 km grids in the DWHC model. For the same reason, one day is regarded as the unit of the model's time step.

Input variables

Meteorological variables

The input meteorological variables include daily precipitation, daily averaged air temperature and daily potential evaporation. The potential evaporation could be the pan evaporation data (E601), or be latent heat output from MM5.

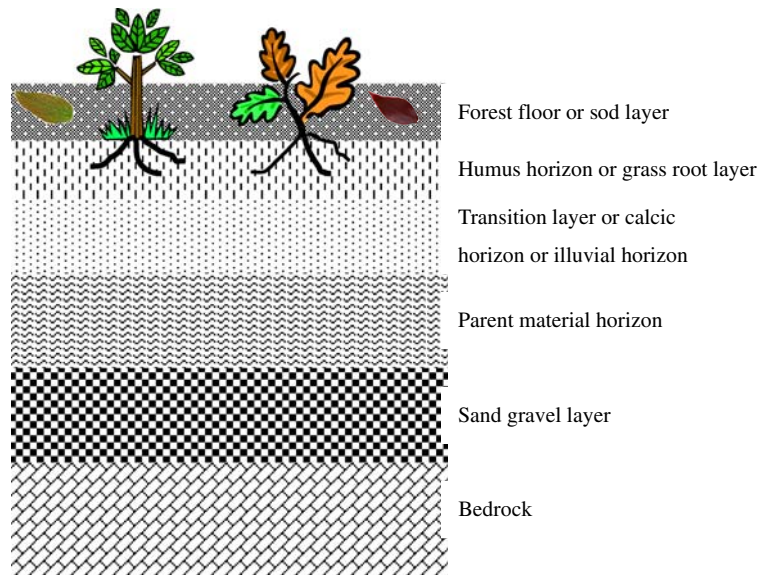
Vegetation information

The used vegetation data include type, coverage, leaf area index (LAI) and saturated interception capacity. According to the latest vegetation map (1:1,000,000, Hou 2001), there are mainly 20 kinds of vegetation landscapes in the research watershed. To simplify, judged by vegetation height and genus, the vegetation in the Heihe mountainous river basin is divided into nine types. With reference to the field data measured in the research watershed (Kang et al. 2005) and in other regions (of same or similar vegetation type. Du et al. 2001; Wang et al. 2004), the LAI and saturated interception capacity of the nine vegetation types are estimated accordingly such as the ones in April 1, June 1 and August 1, etc. Based on the Landsat Thematic Mapper (TM) data shot in July 2000, totally 11 landuse types are differentiated, and are arbitrarily assumed that the vegetation coverage of each landuse type is unchangeable along the time (Chen et al. 2003).

Soil data

In the Heihe mountainous watershed, there are mainly 18 soil types (Chen et al. 2003). Subject to the measured soil profiles (Chen and Qu 1992; Soil Reconnaissance Office of Gansu Province 1993a, b; Chen and Xiao 2003), the soil profile in the research watershed could be divided into 3–5 layers (Fig. 2). The soil parameters required by the DWHC model are estimated by combing the measured results from these soil profiles and from other regions (of same soil type, Liu and Xu 2003a, b; Xu and Liu 2003; Zhang et al. 2001; Zhu et al. 2003; Zhu 1983) together. These soil parameters of each soil layer include depth, porosity, water content at field capacity and wilting point, residual water content, saturated hydraulic conductivity, specific heat of soil grains and characteristic curve of soil moisture. If these

Fig. 2 Soil profiles in mountains watershed of Heihe river basin



data are unavailable, the SOILVISION software (<http://www.soilvision.com>) could be employed just with the soil size as input.

Topography data

The topography data inputs, collected in 1 km grid, are DTM deriving from 1:100,000 topographic map, and its reproduction data from ArcGis software, such as slope, hillside aspect and flow direction data.

Model principles

Solid and liquid precipitation separation and correction

The critical air temperature method (Kang and Ohmura 1994; Kang et al. 1999) is used to separate daily precipitation into rainfall and snowfall.

Because the observed precipitation is less than the factual one due to the air dynamic loss, evaporation loss and wetness loss (Yang et al. 1999), the measured precipitation data should be corrected systemically. According to the experiment results in Urumchi river basin and in Heihe river basin, the corrected factor is about 1.07 and 1.15 for the rainfall and snowfall, respectively.

Canopy interception model

With reference to the methods from SWAT (Arnold et al. 1998), the following canopy interception model is designed:

$$V_{p0} = V_{pmax} \frac{LAI}{LAI_{max}} \tag{1}$$

The LAI in the certain day, as a model input, is used to estimate the everyday value in a year according to the polynomial methods (Arnold et al. 1998).

When one precipitation process starts to run, the canopy storage amount $V_{p,total}$ (mmH₂O) is:

$$V_{p,total} = \begin{cases} V_{p0} & P \geq V_{p0} \\ V_{p0} & (P < V_{p0}) \cap (P + V_{sto} \geq V_{p0}) \\ P + V_{sto} & (P < V_{p0}) \cap (P + V_{sto} < V_{p0}) \end{cases} \tag{2}$$

The precipitation that is intercepted by vegetation canopy V_p (mmH₂O) can be estimated as:

$$V_p = \begin{cases} P & V_{p0} - V_{sto} \geq P \\ V_{p0} - V_{sto} & V_{p0} - V_{sto} < P \end{cases} \tag{3}$$

$$P_{ground} = P(1 - V_{cov}) + (P - V_p) \tag{4}$$

The interception water would be consumed through evaporation process, and at this time the V_{sto} could be calculated by the following equation.

$$V_{sto} = \begin{cases} V_{p,total} - E_0 & V_{p,total} \geq E_0 \\ 0 & V_{p,total} < E_0 \end{cases} \tag{5}$$

Seasonal snow and glacier melting model

The degree-day method (Kang et al. 1999; Chen et al. 2003) is used to quantify both the seasonal snow and glacier melting process. The snowfall down to the glacier

surface is treated as accumulation. Because the glacier in Northwest China is continental, the glacier melting process only occurred from above to below (Chen et al. 2003).

Soil water–heat coupled model

The methods and equations of soil water–heat coupled model are mainly derived from COUP (Jansson and Karlberg 2001), but are simplified for large watershed. The state equation for temperature distribution in the soil matrix, considering the convective heat transfer by liquid and the latent heat transfer by vapor for a layer of freezing soil, is the same as that of SHAW (Flerchinger and Saxton 1989) and COUP (Jansson and Karlberg 2001). Based on the COUP, the continuous equation for heat transfer process in soil can be given numerical solutions according to soil freezing status and some empirical formulas.

Soil freezing status

The soil freezing status can be divided into three types: (1) Unfrozen: soil temperature T_s is not lower than zero; (2) Frozen: T_s is not higher than a threshold temperature value T_f , below which the soil is assumed to be completely frozen except a residual unfrozen amount; (3) Sub-frozen: T_s is higher than T_f but lower than zero. In COUP, the default T_f value is -5°C . However, according to the half-hourly data measured by four Environmental monitoring systems in Heihe river basin (Chen et al. 2006b), here the T_f value is revised as -8°C .

Hydraulic conductivity

The saturated hydraulic conductivity k_0 of unfrozen soil is model input. When soil temperature T_s is lower than zero, k_0 can be corrected as following (Jansson and Karlberg 2001):

$$k'_0 = \begin{cases} k_0 & T_s > 0 \\ k_0(0.54 + 0.023T_s) & T_f \leq T_s \leq 0 \\ 0 & T_s < T_f \end{cases} \quad (6)$$

The unsaturated hydraulic conductivity then can be estimated as (Mualem 1976):

$$k_w = \begin{cases} k_{mat} \left(\frac{\psi_a}{\psi}\right)^{2+(2+n)\lambda} & \theta \leq \theta_m \\ 10 \left(\log \left(k_{mat} \left(\frac{\psi_a}{\psi}\right)^{2+(2+n)\lambda} \right) + \frac{\theta - \theta_s + \theta_m}{\theta_m} \log \left(\left(\frac{k'_0}{k_{mat} \left(\frac{\psi_a}{\psi}\right)^{2+(2+n)\lambda}} \right) \right) \right) & \theta > \theta_m \end{cases} \quad (7)$$

Heat capacity and thermal conductivity

Soil heat capacity C equals the sum of the heat capacities of soil constituents:

$$C = \begin{cases} C_i \rho_i (\theta - \theta_r) + C_l \rho_l \theta_r + C_s \rho_G (1 - \theta_s) & T_s < T_f \\ C_i \rho_i \theta_i + C_l \rho_l \theta_l + C_s \rho_G (1 - \theta_s) & T_f \leq T_s \leq 0 \\ C_l \rho_l \theta + C_s \rho_G (1 - \theta_s) & T_s > 0 \end{cases} \quad (8)$$

Thermal conductivity is a complex function of soil solids and soil moisture:

1. Unfrozen soil

For humus, i.e., organic matter, the thermal conductivity function is adapted from a figure in de Vries (1975):

$$k_{ho} = h_1 + h_2 \theta \quad (9)$$

For unfrozen mineral soil an empirical conductivity function k_{hm} ($\text{J s}^{-1} \text{ m}^{-1} \text{ }^\circ\text{C}^{-1}$) is made out from Kersten (1949):

$$k_{hm} = 0.343 \left(a_1 \log \left(\frac{\theta}{\rho_d} \right) + a_2 \right) 10^{a_3 \rho_d} \quad (10)$$

2. Fully frozen soil

Thermal conductivity of a fully frozen organic soil $k_{ho,i}$ ($\text{J s}^{-1} \text{ m}^{-1} \text{ }^\circ\text{C}^{-1}$) is calculated with a similar equation as for unfrozen organic soils but including a second degree coefficient to account for the influence of ice on the conduction in the soil (de Vries 1975).

$$k_{ho,i} = \left(1 + h_3 Q \left(\frac{\theta}{100} \right)^2 \right) k_{ho} \quad (11)$$

$$Q = - \frac{Q_{total,if} - H}{L_f W_{ice}} \quad (12)$$

Thermal conductivity of fully frozen mineral soil $k_{hm,i}$ is also adapted from Kersten (1949):

$$k_{hm,i} = b_1 10^{b_2 \rho_d} + b_3 \left(\frac{\theta}{\rho_d}\right) 10^{b_4 \rho_d} \tag{13}$$

3. Sub-frozen soil

For temperatures between 0°C and T_f a weighted conductivity is used (Jansson and Karlberg 2001):

$$k_{h,s} = Qk_{h,i} + (1 - Q)k_h \tag{14}$$

Soil total heat content

The heat content of soil at the temperature T_f , $Q_{total,tf}$, is a function of latent and sensitive heat (Jansson and Karlberg 2001):

$$Q_{total,tf} = zC \cdot T_f - L_f w_{ice} \tag{15}$$

$$w_{ice} = w - z\theta_r \rho_l \tag{16}$$

The soil residual volume water content θ_r , is model input, and can be derived from wilting water content θ_{wilt} :

$$\theta_r = d_1 \theta_{wilt} \tag{17}$$

For sub-frozen or unfrozen soil, the heat content function Q_{total} can be given as (Jansson and Karlberg 2001):

$$Q_{total} = L_f w_{ice} \left(\frac{T_s}{T_f}\right)^{\left(\frac{\lambda d_3 + d_2}{d_2 d_3}\right)} + zC_i \cdot T_s \tag{18}$$

Sensible and latent heat component

Soil temperature is the driving force for energy flux in the soil profile. When the temperature drops below 0°C the energy storage in the soil is changed such that liquid water is converted to ice, i.e. change of latent heat, and simultaneous with the temperature decrease, i.e. change of sensitive heat, and vice versa. The latent heat of freezing or thawing is zero when the soil is completely unfrozen or fully frozen.

The temperature in the middle part of a soil layer stands for its average temperature. The heat flow between two adjacent soil layers I and II can be quantified according to the following nine possible cases:

1. Both soil layers I and II are unfrozen

$$H_{I,II} = (k_{h,I} \frac{z_I}{2} + k_{h,II} \frac{z_{II}}{2}) \frac{T_{s,I} - T_{s,II}}{0.5z_I + 0.5z_{II}} \tag{19}$$

Where $H_{I,II}$ is sensible heat flow between soil layer I and II, and k_h is thermal conductivity.

2. Soil layer I is unfrozen while soil layer II is sub-frozen (soil thawing process)

$$f_{lat,II} = L_f w_{ice,II} / Q_{total,tf}^{II} \tag{20}$$

$$r_{II} = \left(1 - Q_{total}^{II} / Q_{total,tf}^{II}\right)^{d_2 \lambda_{II} + d_3} \min \left(1, \left(Q_{total,tf}^{II} - Q_{total}^{II}\right) / \left(Q_{total,tf}^{II} + L_f w_{ice,II}\right)\right) \tag{21}$$

$$H_{I,II} = Q_{total}^{II} (1 - f_{lat,II}) (1 - r_{II}) \tag{22}$$

3. Soil layer I is unfrozen while soil layer II is fully frozen (rarely occurs in soil thawing process): Sensible flow is calculated by Eqs. (19), meanwhile soil layer II's thermal conductivity is outputs of Eqs. (11) or Eqs. (13).
4. Soil layer I is sub-frozen while soil layer II is unfrozen (soil freezing process): similar to case 3.
5. Both the two soil layers are sub-frozen (thawing or freezing process): according to Eqs. (20) to Eqs. (22).
6. Soil layer I is sub-frozen while soil layer II is fully frozen (soil thawing process): similar to case 3.
7. Soil layer I is fully frozen while soil layer II is unfrozen (impossible): similar to case 3.
8. Soil layer I is fully frozen while soil layer II is sub-frozen (soil freezing process): similar to case 2.
9. Both the two soil layers are fully frozen: according to Eqs. (19).

Soil temperature

A change in sensible heat content in the soil, H , results in a new soil temperature, which in turn makes rise to the energy flux that affects the energy storage and so forth. Thus the soil temperature is a function of the sensible heat:

$$T_{s,change}^{II} = \frac{H_{I,II} - H_{II,III}}{C_{II} \rho_{s,II} z_{II}} \tag{23}$$

Once the surface temperature T_0 and initial T_s of each soil layer are input to the model, the T_s in the following days of each soil layer could be deduced. If the T_0 data are absent, it can be estimated by multiplying an empirical constant with the air temperature T (Jansson and Karlberg 2001).

Water content component

After the soil heat starts to transfer and gets a new soil temperature, the liquid and solid water content is estimated accordingly at that time. If the soil layer is

unfrozen, the total water content is equal to the liquid water component; If the soil layer is fully frozen, the liquid water component is the residual water content; if the soil layer is sub-frozen, the change of the solid water content should be estimated:

$$\theta_{i,\text{change}} = \frac{\Delta Q_{\text{total}}}{L_f z} \tag{24}$$

Infiltration model

Two different soil hydraulic properties, water retention curve and unsaturated conductivity function, need to be determined for the water transfer process. The classical characteristic curve of soil moisture can be divided into three segments (Fig. 3, Jansson and Karlberg 2001).

For unfrozen soil layer, the water tension ψ can be estimated as following (Jansson and Karlberg 2001):

$$\frac{\log\left(\frac{\psi}{\psi_x}\right)}{\log\left(\frac{\psi_{\text{wilt}}}{\psi_x}\right)} = \frac{\theta_x - \theta}{\theta_x - \theta_{\text{wilt}}} \quad \theta_{\text{wilt}} < \theta < \theta_x \tag{25}$$

$$\left(\frac{\psi}{\psi_a}\right)^{-\lambda} = \frac{\theta - \theta_r}{\theta_s - \theta_r} \quad \theta_x < \theta < \theta_m \tag{26}$$

$$\psi = \psi_{\text{mat}} - \frac{\theta - \theta_s + \theta_m}{\theta_m} \psi_{\text{mat}} \quad \theta_m < \theta < \theta_s \tag{27}$$

For sub-frozen or fully frozen soil layer, the water tension ψ estimated through Eqs. (25)–(27) should be corrected according to soil water content (Jansson and Karlberg 2001):

$$\psi' = \psi + 200 \frac{\theta_i}{\theta_i + \theta_s - \theta} z \cdot g \tag{28}$$

If the water tension of soil layer I is lower than that of soil layer II, the liquid water will infiltrate from soil layer I to II according to Darcy’s law:

$$q_{I,II} = \begin{cases} \max\left[-k_{w,I} \left(\frac{\psi_I - \psi_{II}}{z_I} - 1\right), (\theta_{I,I} - (\theta_{c,I} - \theta_{I,II}))z_I\right] & \theta_{I,I} \leq \theta_{c,I} - \theta_{I,I} \\ 0 & \theta_{I,I} > \theta_{c,I} - \theta_{I,I} \end{cases} \tag{29}$$

Otherwise, the liquid capillary water will penetrate from soil layer II to I:

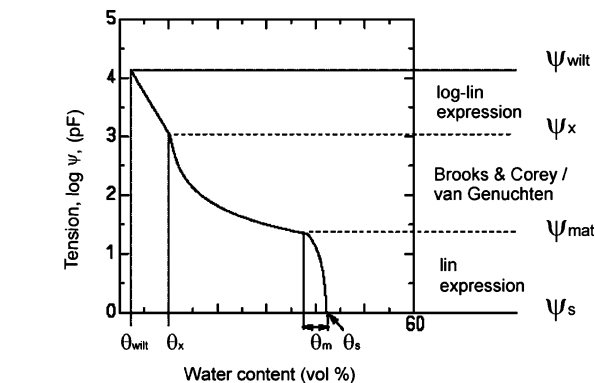


Fig. 3 An example of how three different expressions in the water retention curve are used in different ranges. The pF value corresponds to the logarithm of tension expressed in centimeter (Jansson and Karlberg 2001)

$$q_{I,II} = \min\left[-k_{w,II} \left(\frac{\psi_I - \psi_{II}}{z_{II}} - 1\right), (\theta_{I,II} - \theta_{\text{wilt},II})z_{II}\right] \tag{30}$$

Evapotranspiration model

Due to lack of detailed vegetation information and water flux data, a simplified method is designed to estimate the evapotranspiration, whose process is assumed to occur only in the first soil layers.

Soil evaporation is the function of potential evaporation and effective liquid water content:

$$E'_s = aE_0(\theta_{I,I} - \theta_{r,I}) \tag{31}$$

$$E_s = \min[E'_s, \max(0, (\theta_{I,I} - \theta_{r,I})z_I)] \tag{32}$$

Vegetation transpiration can be estimated by combing potential evaporation, interception evaporation, leaf area index, liquid water content and water content at wilting point together:

$$E'_v = b(E_0 - E_c)(\theta_{I,I} - \theta_{\text{wilt},I})LAI \tag{33}$$

$$E_v = \min[E'_v, \max(0, (\theta_{I,I} - \theta_{\text{wilt},I})z_I)] \tag{34}$$

The average evapotranspiration at a DTM grid is a function of vegetation coverage:

$$E = E_s(1 - V_{cov}) + E_v V_{cov} \quad (35)$$

Runoff generation process

Surface runoff: If the sum of the net liquid water down to ground surface P_{ground} plus snow and glacier meltwater is higher than the saturated hydraulic conductivity k'_0 corrected by soil temperature, surface runoff will appear:

$$R_0 = \max(0, P_{ground} + R_{snow} + R_{glacier} - k'_0) \quad (36)$$

The surplus liquid water after surface runoff process is assumed to infiltrate into the first soil layer completely. Then, the liquid and the solid water contents within the first soil layer are calculated to decide whether the subsurface water appears or not:

$$R_l = \max \left[0, \begin{pmatrix} 0 & \theta_{l,I} \leq \theta_{r,I} \\ (\theta_{l,I} - (\theta_{s,I} - \theta_{i,I}))z_I & \theta_{l,I} > \theta_{r,I} \end{pmatrix} \right] \quad (37)$$

Now the heat and water transfer processes in soil should be estimated as following:

1. Estimate C (Eq. (8)), w_{ice} (Eq. (16)) and $Q_{total,tf}$ (Eq. (15)) within the first soil layer;
2. Determine their freezing status according to T_s of the first soil layer and the second soil layer;
3. Estimate the water tension of both the first and the second soil layer (Eqs. (25)–(28));
4. Calculate the unsaturated hydraulic conductivity according to their freezing status (Eqs. (6) and (7));
5. Estimate infiltration or capillary water between the first and second soil layers (Eqs. (29) and (30));
6. Quantify the liquid and solid water content in the first soil layer after runoff's generation (Eq. (37)) and infiltration processes;
- 7) Compute the actual evapotranspiration from the first soil layer (Eqs. (31)–(35));
8. Estimate the C and thermal conductivity of the first soil layer according to its freezing status and water content (Eqs. (9)–(14));
9. Calculate the sensible heat flow between soil surface and the first soil layer;
10. Estimate the heat variables of the second soil layer (Eqs. (9)–(18));
11. Quantify the sensible heat flow between the first and second soil layers (Eqs. (19)–(22));
12. Estimate the change of soil temperature within the first soil layer (Eq. (23));

13. Calculate the change of water content within the first soil layer (Eq. (24)).

Here ends up the water and heat cycle within the first soil layer, as for that of the rest soil layers, it can be deduced by being analogous with that above.

Flow concentration model

Flow concentration process is very complex. Here, a simplified method similar to isochrones is used, which uses topographic data and flow direction data as inputs. The flow direction data are derived from DTM according to the Deterministic eight-neighbors (D8) principle (Fairfield and Leymarie 1991) by using Arcgis software. The flow accumulation time between two adjacent DTM grids i and j can be estimated as:

$$t_{i,j} = \frac{l_i}{m_1 \tan(\beta_i)^{m_2}} + \frac{l_j}{m_1 \tan(\beta_j)^{m_2}} \quad (38)$$

The differences of the flow accumulation time $t_{i,j}$ between surface flow and subsurface flow from different soil layers are controlled by parameters m_1 and m_2 at a DTM grid. Here, parameter m_1 or m_2 is different among different soil layers, while maintaining the same for the same soil layer, for example, the m_1 or m_2 value for soil layer II is assumed unchangeable in the whole watershed.

According to the flow direction data, the total flow time can be estimated from any DTM grid to the outlet of the research watershed (Eq. (38)). Based on the flow time data, the total runoff of the research watershed in a certain date then can be estimated as the sum of the runoff from land surface and all the soil layers that arrives to the outlet of the research in the same date.

Model parameters

Although the DWHC model is somewhat physically based, it still has some conceptual parameters due to its practical difficulty in collecting enough soil, hydrometeorological and detailed topography data in large watershed area. These parameters can be divided into two types:

1. Empirical constants: these constants, with some physical basement, are mainly derived either from relative field measurements or from statistical results. They are unadjusted during the model calibration processes. These constants include thresholds of air temperature defining precipitation as liquid or solid components (Chen et al. 2003), corrected coefficients for observed liquid and solid precipitation (Yang et al. 1999), degree-day factors for snow or glacier melting

processes ($4.5 \text{ mm}^\circ\text{C}^{-1}\text{d}^{-1}$ for snow and $5.0 \text{ mm}^\circ\text{C}^{-1}\text{d}^{-1}$ for glacier; Wang 2004), the threshold soil temperature T_f below which the soil is assumed to be fully frozen and base flow constant. The base flow constant is close to the measured daily runoff in late winter. The final simulated daily runoff of the research watershed is the sum of the base flow, surface flow and subsurface flow. The reason that the base flow constant appears is that the model has not considered the effects of the deep groundwater, which should be improved in the future.

- Adjustable parameters: these parameters should be adjusted in the model calibration processes, which include two parameters for evapotranspiration processes (Eqs. (33)–(35)) and two flow time coefficients (Eq. (38)).

In addition, the initial value of certain variables including soil temperature and soil water content should be given before simulation. The initial values are derived from measured data at meteorological stations and automatic stations (Chen et al. 2006b) and from statistical results between air temperature and soil temperature.

Summary

Frozen ground impacts to water cycle is little considered in the distributed hydrological models for a watershed in a large scale. Be that as it might, in the watershed in cold regions, the water and heat transfer process is dominant, evidently affecting water cycle processes and environments, under global warming today, especially in the regions with arid climate and weak ecology, such as in the Northwest China. The DHWC model in this essay is an initial attempt to combine water and heat coupled processes with many methods from SHAW and COUP. However, the model is relatively simple with a few meteorological variables as inputs due to hard practical conditions—with rare meteorological stations in the large watershed. Thus, the model still has some conceptual parameters, and need to be improved in the future.

With 1 km DTM grids in daily scale, the DWHC model presents the basic hydrological processes in Heihe mountainous watershed by dividing soil into 3–5 layers for each of 18 soil types, 9 vegetation types and 11 landuse types in accordance with the previous research results. The model can compute the continuous equation of heat and water flow in the soil and can estimate them continuously, by numerical methods or some empirical formula, which depends on freezing soil status.

This paper describes only the model structure and basic equations, while in the next papers, the model calibrating

results using the data measured at meteorological stations, as well as MM5 outputs, will be introduced.

Acknowledgments This work has been mainly supported by Chinese National Sciences Foundation Committee and Chinese Academy of Sciences (KZCX2-YW-301-3 and 40401012). The authors would also thank Dr. Jason Yan for English improvements.

References

- Arnold JG, R Srinivasin, RS Muttiah, JR Williams (1998) Large area hydrologic modeling and assessment: Part I. Model development. *JAWRA* 34(1):73–89
- Bayard D, Stähli M (2005) Effects of frozen soil on the groundwater recharge in Alpine areas. In: de Jong C, Collins D, Ranzi R (eds) *Climate and hydrology in mountain areas*. Wiley, Chichester, pp 73–83
- Chen LH, Qu YG (1992) Water and land resources and their rational development and utilization in the Hexi region (in Chinese). Science Press, Beijing
- Chen LH, Xiao HL (2003) Mountainous soil and its utilization in Hexi region (in Chinese). Ocean Press, Beijing
- Chen RS, Kang ES, Yang JP, Zhang JS (2003) A distributed runoff model for inland river mountainous basin of northwest china. *J Geogr Sci* 13(3):363–372
- Chen RS, Kang ES, Ji XB, Yang JP, Yang Y (2006a) Cold regions in China. *Cold Reg Sci Technol* 45:95–102
- Chen RS, Kang ES, Ji XB, Yang JP, Zhang ZH, Yang Y (2006b) Spatial and temporal variations of atmospheric CO_2 concentration and its response to meteorological variables in Heihe river basin, Northwest China. *J Environ Sci (China)* 18(4):708–715
- Du ZC, Yang ZG, Cui XY (2001) A comparative study on leaf area index of five plant communities in typical steppe region of Inner Mongolia (in Chinese). *Grassl China* 23(5):13–18
- Fairfield J, Leymarie P (1991) Drainage network from grid digital elevation models. *Water Resour Res* 27(4):29–61
- Flerchinger GN, Saxton KE (1989) Simulation heat and water model of a freezing snow-residue-soil system I. Theory and development. *Trans ASAE* 32(2):565–571
- Gao YH, Lu SH, Cheng GD (2004) Simulation of rainfall-runoff and watershed convergence process in the upper reaches of Heihe river basin, July 2002. *Sci China Ser D* 47(suppl 1):1–8
- Harlan RL (1973) Analysis of coupled heat-fluid transport in partially frozen soil. *Water Resour Res* 9:1314–1323
- Hou XY (2001) *Vegetation atlas in China (1:1000000)* (in Chinese). Science Press, Beijing
- Ippisch O (2001) *Coupled transport in natural porous media*. PhD Thesis, University of Heidelberg, Heidelberg, Germany
- Jansson P-E, Halldin S (1979) Model for the annual water and heat flow in a layered soil. In: Halldin S (ed) *Comparison of forest and energy exchange models*. International Society for ecological modelling, Copenhagen, Denmark, pp 145–163
- Jansson P-E, Karlberg L (2001) *Coupled heat and mass transfer model for soil-plant-atmosphere systems*. TRITAAMI Report 3087, ISSN 1400–1306, KTH Stockholm, Sweden
- Kang ES, Ohmura A (1994) Energy, water amount and runoff balance model in a glacier watershed, Tien Shan (in Chinese). *Sci China Ser B* 24(9):983–991
- Kang ES, Cheng GD, Lan YC, Jin HJ (1999) A model for simulating the response of runoff from the mountainous watershed of inland river basins in the arid area of northwest China to climatic changes. *Sci China Ser D* 42(suppl):52–63

- Kang ES, Cheng GD, Song KC, Jin BW, Liu XD, Wang JY (2005) Simulation of energy and water balance in soil-vegetation-atmosphere transfer system in the mountain area of Heihe River Basin at Hexi Corridor of northwest China. *Sci China Ser D* 48(4):538–548
- Kersten MS (1949) Thermal properties of soils. Inst. of Technology, Eng. Exp. Station, Bull. No. 28, Univ. Minnesota, Minneapolis
- Liu JL, Xu SH (2003a) Figuring soil water characteristic curve based on particle size distribution data: application of fractal models (in Chinese). *Acta Pedologica Sinica* 40(1):46–52
- Liu JL, Xu SH (2003b) Estimation of soil water retention characteristics with a non-similar-media method (in Chinese). *J Hydraul Eng* 4:80–84
- Mualem Y (1976) A new model for predicting the hydraulic conductivity of unsaturated porous media. *Water Resour Res* 12:513–522
- Soil Reconnaissance Office of Gansu Province (1993a) Gansu soil (in Chinese). Chinese Agriculture Press, Beijing
- Soil Reconnaissance Office of Gansu Province (1993b) Gansu soil type (in Chinese). Gansu Science and Technology Press, Lanzhou
- Stähli M, Nyberg L, Mellander P-E, Janson P-E, Bishop KH (2001) Soil frost effects on soil water and runoff dynamics along a boreal transect: 2. Simulations. *Hydrol Process* 15:927–941
- de Vries DA (1975) Heat transfer in soils. In: de Vries DA, Afgan NH (eds) Heat and mass transfer in the biosphere. I. Transfer processes in plant environment. Scripta Book Co., Washington DC, pp 5–28
- Wang J (2004) Modeling the runoff processes in a cold grassland at the Atuoyinake headstream, southwest Tianshan Mountains, China (in Chinese). Master Thesis, CAREERI, Lanzhou, China
- Wang XP, Kang ES, Zhang JG, Li XR (2004) Comparison of interception loss in Shrubby and Subshrubby Communities in the Tengger Desert of Northwest China (in Chinese). *J Glaciol Geocryol* 26(1):89–94
- Xu SH, Liu JL (2003) Fractal approach for estimating soil water retention curves of various textures (in Chinese). *J Hydraul Eng* 1:78–82
- Yang DQ, Goodison BE, Metcalfe JR, Louie P, Leavesley G, Emerson D, Hanson CL, Golubev VS, Elomaa E, Gunther T, Prunburn T, Kang ES, Milkovic J (1999) Quantification of precipitation measurement discontinuity induced by wind shields on national gauges. *Water Resour Res* 35(2):491–508
- Yang QH, Snyder JP, Tobler WR (2000a) Map projection transformation: principles and applications. Taylor & Francis, London
- Yang ZN, Liu XR, Zeng QZ, Chen ZT (2000b) Hydrology in cold regions of China (in Chinese). Science Press, Beijing
- Zhang XY, Zhang L, Liu CM (2001) On describing the hydraulic properties of unsaturated soil in piedmont of Mt. Taihang (in Chinese). *Acta Agriculturae Boreali Sinica* 16(2):75–82
- Zhu ZX (1983) Soil sciences (in Chinese). Chinese Agriculture Press, Beijing
- Zhu AN, Zhang JB, Cheng ZH (2003) A simple method to estimate water retention curves of light-textured soil (in Chinese). *Chin J Soil Sci* 34(4):253–258