Modification and Comparison of Thermal and Hydrological Parameterization Schemes for Different Underlying Surfaces on the Tibetan Plateau in the Warm Season

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Abstract In this paper, three parameterization schemes, one considering gravel influence (test1), one considering organic carbon influence (test2), and one comprehensively considering gravel and organic carbon influence (test3), were set up to modify different typical underlying surfaces of the Tibetan Plateau (TP). In addition, their soil thermal properties and hydraulic property variations were discussed. Additionally, discussing the key thermal and hydraulic parameters affected the performance of different schemes from the perspective of observation data in the TP to improve the simulation ability of soil temperature and soil moisture in the plateau areas. Compared the original Community Land Model (CLM) scheme, test1 resulted in higher soil temperature and lower soil moisture, while test2 had lower soil temperature and higher soil moisture. The key thermal and hydraulic parameters are the changes in the saturated thermal conductivity and the thermal conductivity of dry soil and the variations of the porosity and exponent B, respectively. The test3 scheme was the same as test2 for the modified changes of thermal properties, except that the proportion of change was slightly different. In terms of soil thermal properties, test2 and test3 were better at 0–20 cm depth, while test1 and test3 were better for the deeper (40 cm) simulation. Regarding hydraulic properties, the test1 and test3 schemes performed better on the Gobi and alpine meadows at 20–40 cm depth, while the original CLM scheme and test2 performed better on the underlying grassland surface. Test3 was better at balancing the relationship between the thermal and hydraulic parameters and could be used for the further research on the entire plateau area.

1. Introduction

Known as the “third pole” of the earth (Qiu, 2008), the Tibetan Plateau (TP) has a vast area and an average altitude of over 4,000 m. It is the highest plateau in the world and one of the most sensitive areas in terms of its response to global climate change (Wang et al., 2015). Studies have shown that the energy and water cycle processes of the TP play an important role in climate and environmental changes in China and even globally (Wu et al., 1999; Ye, 1981). Soil moisture affects the climate by changing the sensible heat, latent heat and long-wave radiation flux from the surface to the atmosphere. Changes in climate also affects the thermodynamic properties and hydrological processes of the soil itself, causing changes in various parameters on the surface. Further influence on climate change generally has an impact on short-term climate change; conversely, climate change can cause changes in soil moisture, and this impact is on interannu-al and interdecadal scales. This shows that climate change and soil moisture interact and influence each other. Araghi et al. (2019) used an ensemble of 17 General Circulation Models (GCMs) from the CMIP5 to estimate future soil temperature in northeast Iran. The research shows increased soil temperature can create faster and higher water deficits in soil, especially in arid to semi-arid climates. The plateau surface is composed of many different underlying surface types, such as land, water, ice and snow. The underlying surface affects climate change through the exchange of sensible heat, latent heat and radiant energy with the atmosphere. Due to its different vegetation types, complex surface coverage, topography and soil texture, the water, heat and other parameters of the plateau land are more complex and changeable than the
other underlying surfaces, resulting in different climate changes. Therefore, it is very important to study the land surface processes of the different underlying surfaces of the plateau (Dong et al., 2000; Luo et al., 1995; Wang et al., 2003; G. X. Wu et al., 1998; T. W. Wu et al., 2003).

Field scientific experiments have shown that the gravel content of the TP is very high and widely distributed (Arocena et al., 2012; Luo et al., 2009; Ohtsuka et al., 2008; X. D. Wu et al., 2012). However, in the existing hydrothermal parameterization schemes, the soil texture considers only the water transport and heat conduction properties of fine soil with a particle diameter of less than 2 mm and ignores the impact of gravel with a particle diameter greater than 2 mm on the soil hydrothermal properties (Brouwer & Anderson, 2000). The hydrothermal properties of gravel are completely different from those of fine soil. Gravel can change soil porosity and increase the tortuosity of soil water conduction paths, thereby affecting soil water transmission and solute transport. It leads to changes in soil hydraulic conductivity and the soil infiltration rate and affects the soil water content (Childs et al., 1990; C. L. Ma, Lü & Pan, 2020; Y. Ma, Hu, et al., 2020; Mehuys et al., 1975). The soil water content affects the atmospheric boundary layer through sensible heat and latent heat, which in turn affect precipitation (Luan et al., 2018; Y. S. Ma et al., 2019). Studies have found that the thermal diffusivity of gravel soil is higher than that of fine soil, and diurnal changes in temperature will also be reflected in gravel soil (Li et al., 2002). Soil with a larger gravel content has higher thermal conductivity in the frozen state (He et al., 2017). Pan et al. (2015) developed a soil parameterization scheme that was more suitable for the soil characteristics of the TP based on the high gravel content and effectively improved the simulation of Community Land Model v4.0 (CLM4.0) soil water and heat properties.

The TP has a lower temperature than areas at the same latitude; it also has slower decomposition of organic matter, which leads to the accumulation of organic matter in the surface soil. In addition, the accumulation of organic carbon on the underlying surface of plateau meadows and grasslands is dominant (Wang et al., 2002). The water and heat transfer characteristics of soil organic carbon are quite different from those of minerals. Organic carbon can change the soil water content, water characteristic curve, saturated hydraulic conductivity and other soil hydrodynamic parameters and increase the effective soil water holding capacity (Q. Ma et al., 2014; Shan et al., 1998), thereby enhancing the ability of soil to absorb heat and increasing soil temperature (Lawrence & Slater, 2008). Luo et al. (2008) found that the Farouki scheme used in CoLM simulation often overestimated the thermal conductivity of alpine grassland soils. The study found that the high content of organic carbon in the surface soil of alpine grassland resulted in a significant decrease in soil thermal conductivity and heat capacity (Yang et al., 2005) and further confirmed that the impact of organic carbon on soil hydrothermal properties should be considered in the model (Yang et al., 2009).

However, the existing models still cannot accurately reflect the land surface and hydrological processes of heterogeneous underlying surfaces, and there is insufficient research on the characteristics of the land surface process and hydrothermal parameters of different typical underlying surfaces in the TP and their impact on climate change. The parameterization scheme only considering gravel influence was not perfect about the soil heat (C. L. Ma, Lü, & Pan, 2020; Y. Ma, Hu, et al., 2020), while it mainly considering organic carbon influence was not perfect about the soil water (Chen et al., 2012; Q. Ma et al., 2014). There is still a lack of parameterization schemes suitable for different typical underlying surfaces (Wen et al., 2011).

This study investigates the applicability of different thermal and hydrological parameterization schemes for different underlying surfaces on the TP in the warm season and discusses the changes in the hydrothermal properties of different underlying surfaces. The main contents of the paper are as follows. Section 2 introduces the data and methods. Section 3 discusses the modification of the thermal and hydraulic parameterization schemes in CLM5.0. Section 4 explores the soil thermal properties of the different underlying surfaces. Section 5 analyses the soil hydraulic properties of the different underlying surfaces. Finally, we end with a summary and a main discussion in Section 6.

2. Data and Methods

The soil temperature/moisture used in this paper were CLM5.0 simulation output driven by in situ observations, including precipitation, air pressure, air temperature, specific humidity, wind speed, solar radiation and longwave radiation information measured every 0.5 hr at QOMS station in 2016 and at BJ and A’rou...
station in 2019. And each station cycled the atmospheric forcing data set to run spin-up for 100 years to make sure the surface variables entered steady state (Yang et al., 1995). As the latest land surface model version released by CESM, CLM5.0 is currently the most promising and advanced land surface process model. Compared with CLM4.5, most of the main components of the CLM5.0 model have been modified. In particular, significant changes have taken place in soil and plant hydrology and carbon and nitrogen cycles. Most of the development focus is on more comprehensively and clearly reflecting the processes of land use and land cover changes. Hydrological improvements include the introduction of dry surface layer-based soil evaporation resistance parameterization (Swenson & Lawrence, 2014) and revised canopy interception parameterization (Brunke et al., 2016; Swenson & Lawrence, 2015). The adaptive time-stepping solution of Richard’s equation was introduced, which improved the accuracy and stability of the numerical solution of soil moisture.

The in situ observations of this study were derived from the Qomolangma Atmospheric and Environmental Observation and Research Station, Chinese Academy of Sciences (CAS) (QOMS), the Nagqu Station of Plateau Climate and Environment, CAS (BJ) (C. L. Ma, Lü, & Pan, 2020; Y. Ma, Hu et al., 2020), and the Qilian Mountains integrated observatory network, National TP Data Center (A’rou) (Liu et al., 2020). Figure 1 depicts geographical distribution of observational sites over the TP, and Table 1 shows the basic information of the three stations, including their locations, land covers, soil types, measured soil organic carbon mass contents ($socEm$) and soil gravel mass proportions ($gEm$) (Chen et al., 2012; C. L. Ma, Lü, & Pan, 2020; Y. Ma, Hu, et al., 2020). Figure 2 shows the change of wind speed and precipitation data with time at QOMS, BJ, and A’rou station. The half-hour precipitation increased after about June 15 and this time is the beginning of the rainy season on the TP. A’rou and BJ station have significantly more precipitation than QOMS station, which is related to the location of the two stations more eastward. And the wind speed change at QOMS station is the largest than the other two stations. All simulations in this paper are the results of replacing the observed percent sand, percent clay and organic matter density at the soil level (Chen et al., 2012; Su et al., 2020). The organic carbon content and gravel content of different plateau underlying surfaces were quite different. The gravel content of the three underlying surfaces increased and the organic carbon content decreased with increasing soil depth. From the QOMS station to the A’rou station, the organic carbon content increased, and the gravel content decreased. Therefore, the three selected stations were very representative of the different typical underlying surfaces of the plateau.
### Table 1
**The Basic Information of the Three Stations**

| Station | Latitude | Longitude | Land cover   | Soil type  | Depth (cm) | $m_{soc}$ (%) | $m_{g}$ (%) |
|---------|----------|-----------|--------------|------------|------------|---------------|-------------|
| QOMS    | 28.36°N  | 86.95°E   | Gobi         | Sand and gravel | 0–10      | 0.84          | 21.5        |
|         |          |           |              |            | 10–20      | 0.42          | 30.61       |
|         |          |           |              |            | 20–30      | 0.36          | 42.5        |
|         |          |           |              |            | 30–40      | 0             | 55          |
| BJ      | 31.37°N  | 91.90°E   | Alpine meadow | Sandy      | 0–10      | 2.1           | 2.1         |
|         |          |           |              |            | 10–20      | 1.3           | 2.3         |
|         |          |           |              |            | 20–30      | 1.1           | 7.9         |
|         |          |           |              |            | 30–40      | 1.5           | 3.2         |
| A’rou   | 38.04°N  | 100.46°E  | Grassland    | Silt loam  | 0–10      | 7.5           | 0           |
|         |          |           |              |            | 10–20      | 5.3           | 0.3         |
|         |          |           |              |            | 20–30      | 3.5           | 0           |
|         |          |           |              |            | 30–40      | 2.8           | 2.1         |

**Figure 2.** The change of wind speed and precipitation data with time at QOMS station (a and b), BJ station (c and d), and A’rou station (e and f).
The soil temperature is an important indicator for testing the simulated performance of the modified soil thermal properties. The change in soil temperature directly affects the energy balance of land-atmosphere interactions and cycles, thus having a feedback effect on climate change (Luo et al., 2009). Therefore, verifying the revised soil temperature is a significant part of the numerical simulation process. This paper mainly used Taylor diagrams to evaluate the pros and cons of different schemes. The bias in the figure is the relative bias (%) and calculated as follows:

$$\text{Bias} = \frac{1}{N} \sum_{n=1}^{N} \left( \frac{\text{test}_n - \text{OBS}_n}{\text{OBS}_n} \right) \times 100,$$

(1)

where test\(_n\) and OBS\(_n\) represent different parameterization schemes and observations, respectively.

\(E\) is the centered root-mean-square difference between the tests and in situ observations, and \(\sigma^2_{\text{test}}\) and \(\sigma^2_{\text{OBS}}\) are the variances of the test and observation, respectively. \(R\) is the test pattern correlation with in situ observations.

$$E^2 = \sigma^2_{\text{test}} + \sigma^2_{\text{OBS}} - 2 \times \sigma_{\text{test}} \times \sigma_{\text{OBS}} \times R,$$

(2)

The use of the \(E\) index aims to quantify the similarity between different schemes and observations in terms of their correlation, standard deviation and the amplitude of their variations (Taylor, 2001) to select a more suitable hydrothermal parameterization scheme.

| Parameter | CLM | Test1 | Test2 | Test3 |
|-----------|-----|-------|-------|-------|
| \(f_{\text{init}}\) | \(\rho_{\text{init}} + \rho_{\text{max}}\) | | | |
| \(\lambda_{\text{dry,init}}\) | 8.80 (%sand) + 2.92 (%clay) | | | |
| \(\lambda_{\text{dry,init}}\) | 0.135(1 - \(\theta_{\text{sat}}\)) | | | |
| \(\theta_{\text{sat,init}}\) | 0.489 - 0.00126 (%sand) | | | |
| \(\Psi_{\text{sat,init}}\) | \(-10.0 \times 10^{1.88-0.013}\) | | | |
| \(k_{\text{sat,init}}\) | 0.0070556 \times 10^{0.884+0.015} | | | |
| \(K_{x,i}\) | | | | |

The soil temperature is an important indicator for testing the simulated performance of the modified soil thermal properties. The change in soil temperature directly affects the energy balance of land-atmosphere interactions and cycles, thus having a feedback effect on climate change (Luo et al., 2009). Therefore, verifying the revised soil temperature is a significant part of the numerical simulation process. This paper mainly used Taylor diagrams to evaluate the pros and cons of different schemes. The bias in the figure is the relative bias (%) and calculated as follows:

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$$E^2 = \sigma^2_{\text{test}} + \sigma^2_{\text{OBS}} - 2 \times \sigma_{\text{test}} \times \sigma_{\text{OBS}} \times R,$$

(2)

The use of the \(E\) index aims to quantify the similarity between different schemes and observations in terms of their correlation, standard deviation and the amplitude of their variations (Taylor, 2001) to select a more suitable hydrothermal parameterization scheme.
3. Modified Thermal and Hydraulic Parameterization Schemes in CLM5.0

In this paper, CLM represents the original thermal and hydraulic parameterization schemes of CLM5.0. The revision scheme of test1 mainly considers the impact of gravel on soil thermal and hydraulic properties (Y. Ma et al., 2020; C. L. Ma, Lü, & Pan, 2020). The revision scheme of test2 mainly considers the impact of soil organic carbon on soil thermal and hydraulic properties (Chen et al., 2012; Q. Ma et al., 2014). In addition, see below for the modification of CLM scheme and test3 in terms of hydrothermal parameterization.

### 3.1. CLM Scheme

#### 3.1.1. Thermal Parameterization of CLM Scheme

The soil temperature of the CLM5.0 scheme adopts the second law of heat conduction in one dimension:

$$
\frac{\partial T}{\partial t} = \frac{1}{\lambda} \frac{\partial}{\partial z} \left( \lambda \frac{\partial T}{\partial z} \right)
$$

Equation 3 shows that soil temperature is a function of soil heat capacity ($J \cdot m^{-3} \cdot K^{-1}$) and thermal conductivity ($W \cdot m^{-1} \cdot K^{-1}$).

The soil layer organic matter fraction $f_{om,i,j}$ is

$$
f_{om,i,j} = \frac{\rho_{om,i}}{\rho_{om,max}}
$$

where $\rho_{om,max} = 130 \text{ kg m}^{-3}$.

Soil thermal conductivity $\lambda_i$ ($W \cdot m^{-1} \cdot K^{-1}$) is from Farouki (1981) and calculated as follows:
where \( \lambda_{s_i} \) is the thermal conductivity of soil solids, calculated as follows:

\[
\lambda_{s_i} = (1 - f_{\text{om},i}) \lambda_{\text{min},i} + f_{\text{om},i} \lambda_{\text{om},i},
\]

where \( \lambda_{\text{min},i} \) is the thermal conductivity of mineral soil, calculated as follows:

\[
\lambda_{\text{min},i} = \frac{8.80 \% \text{ sand} + 2.92 \% \text{ clay}}{(\% \text{ sand}) + (\% \text{ clay})}.
\]

The thermal conductivity of dry soil \( \lambda_{\text{dry},i} \) (W m\(^{-1}\) K\(^{-1}\)) is calculated as follows:

\[
\lambda_{\text{dry},i} = (1 - f_{\text{om},i}) \lambda_{\text{dry},\text{min},i} + f_{\text{om},i} \lambda_{\text{dry},\text{om},i},
\]

where \( \lambda_{\text{dry},\text{min},i} \) is the thermal conductivity of dry mineral soil and is calculated as follows:

\[
\lambda_{\text{dry},\text{min},i} = \frac{0.135 \rho_{d,i} + 64.7}{2700 - 0.947 \rho_{d,i}},
\]

where \( \rho_{d,i} = 2700 (1 - \theta_{\text{sat},i}) \).

Figure 4. The same as Figure 3, but for station BJ.

(a) the thermal conductivity of soil solids

(b) the saturated thermal conductivity

(c) the thermal conductivity of dry soil

(d) soil thermal conductivity

(e) the heat capacity of soil solids
The Kersten number $K_{e,j}$ (Kersten, 1949) is calculated as follows:

$$K_{e,j} = \begin{cases} 
\log(S_{e,j}) + 1 \geq 0 & T_i \geq T_f \\
S_{e,j} & T_i < T_f 
\end{cases},$$

(11)

where $S_{e,j}$ is the wetness of the soil with respect to saturation.

The volumetric heat capacity $c_i$ (J m$^{-3}$ K$^{-1}$) for soil (de Vries, 1963) is as follows:

$$c_i = c_{i,j}(1 - \theta_{sat,j}) + \frac{w_{ice,j} C_{ice}}{\Delta z_i} + \frac{w_{liq,j} C_{liq}}{\Delta z_i},$$

(12)

where $C_{ice}$ and $C_{liq}$ are the specific heat capacities (J kg$^{-1}$ K$^{-1}$) of ice and liquid water, respectively. $c_{i,j}$ is the heat capacity of soil solids, calculated as follows:

$$c_{i,j} = (1 - f_{s,om,j}) c_{i,\text{min},j} + f_{s,om,j} c_{i,\text{min,om}},$$

(13)

where $c_{i,\text{om}} = 2.5 \times 10^6$ J m$^{-3}$ K$^{-1}$, and $c_{i,\text{min,om}}$ is the heat capacity of mineral soil solids, calculated as follows:

$$c_{i,\text{min,om}} = \frac{2.128 (\% \text{sand}) + 2.385 (\% \text{clay})}{(\% \text{sand}) + (\% \text{clay})_i} \times 10^6.$$

(14)

### 3.1.2. Hydraulic Parameterization of CLM Scheme

The change in soil water content (Dingman, 2002) is as follows:

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left[ k \left( \frac{\partial \Psi}{\partial z} + 1 \right) \right].$$

(15)
Equation 17 shows that the soil water content is a function of the hydraulic conductivity $k$ (mm s$^{-1}$) and soil matric potential $\Psi$ (mm).

The porosity is as follows:

$$\theta_{sat,t} = (1 - f_{soc,t}) \theta_{sat,min,t} + f_{soc,t} \theta_{sat,om,t},$$

where the porosity of the mineral soil $\theta_{sat,min,t}$ is as follows:

$$\theta_{sat,min,t} = 0.489 - 0.00126(\%sand)_t.$$  \hspace{1cm} (17)

The exponent $B_t$ is as follows:

$$B_t = (1 - f_{soc,t}) B_{min,t} + f_{soc,t} B_{om,t},$$

where $B_{om} = 2.7$.

The exponent $B$ of the mineral soil is as follows:

$$B_{min,t} = 2.91 + 0.159(\%clay)_t.$$  \hspace{1cm} (19)

The soil matric potential (mm) $\Psi$ is as follows:

$$\Psi_t = \Psi_{sat,t} \left( \frac{\theta_t}{\theta_{sat,t}} \right)^{B_t} \geq -1 \times 10^8,$$ \hspace{1cm} (20)

where the saturated soil matric potential (mm) $\Psi_{sat,t}$ (Farouki, 1981; Letts et al., 2000) is as follows:

$$\Psi_{sat,t} = (1 - f_{soc,t}) \Psi_{sat,min,t} + f_{soc,t} \Psi_{sat,om,t},$$ \hspace{1cm} (21)

where the saturated mineral soil matric potential $\Psi_{sat,min,t}$ (Cousin et al., 2003; Pan et al., 2015) is as follows:

$$\Psi_{sat,min,t} = -10.0 \times 10^{1.88-0.0131(\%sand)_t},$$ \hspace{1cm} (22)

The saturated hydraulic conductivity for mineral soil $k_{sat,om}$ (Cosby et al., 1984; Peck et al., 1979) is as follows:

$$k_{sat,om} = 0.070556 \times 10^{-0.884+0.015(\%sand)_t},$$ \hspace{1cm} (23)

3.2. Test3 Scheme

3.2.1. Thermal Parameterization of Test3 Scheme

Test3 uses the volume percentage of soil organic carbon ($V_{soc}$) to highlight the proportion of soil organic matter (Chen et al., 2012).

$$V_{soc} = \frac{\rho_p (1 - \theta_{sat,min,t}) m_{soc}}{\rho_{soc} (1 - m_{soc}) + \rho_p (1 - \theta_{sat,min,t}) m_{soc} + (1 - \theta_{sat,min,t}) \rho_{soc} m_{x} / (1 - m_{x})},$$ \hspace{1cm} (24)

$\lambda_{s,i}$ is the thermal conductivity of soil solids, calculated as follows:

$$\lambda_{s,i} = (1 - V_{soc}) \lambda_{s,om,i} + V_{soc} \lambda_{s,om},$$ \hspace{1cm} (25)

Figure 6. Variations in soil temperature from QOMS in situ observations (black), CLM (blue), test1 (orange), test2 (green), and test3 (red) at 10, 20, and 40 cm depths from June to August 2016.
where \( \lambda_{s,m} = 0.25 \text{ W m}^{-1} \text{ K}^{-1} \), and \( \lambda_{s,min,i} \) is the mineral soil solid thermal conductivity, calculated as follows (Chen et al., 2012):

\[
\lambda_{s,min,i} = \lambda_{q}^{0.7} \lambda_{soc}^{1.0},
\]

where \( \lambda_{q} = 7.7 \text{ W m}^{-1} \text{ K}^{-1} \), \( \lambda_{soc} = 0.25 \text{ W m}^{-1} \text{ K}^{-1} \), \( \lambda_{0} = 2.0 \text{ W m}^{-1} \text{ K}^{-1} \), and \( q_{0} = 0.5\% \text{ sand}_{\text{new}} \), calculated as follows:

\[
\% \text{ sand}_{\text{new}} = \frac{(1 - V_{soc} - V_{g}) \% \text{ sand} + V_{g}}{1 - V_{soc}},
\]

where \( V_{g} \) is the volumetric content of gravel in soils, calculated as follows:

\[
V_{g} = \frac{\rho_{soc}(1 - \theta_{sat,min,i})m_{g}}{(1 - m_{g})[\rho_{soc}(1 - m_{soc}) + \rho_{p}(1 - \theta_{sat,min,i})m_{soc} + (1 - \theta_{sat,min,i})\rho_{soc}m_{g}]}.
\]

where \( \rho_{soc} = 130k_{g} \text{ m}^{-3} \) is the bulk density of peat, \( \rho_{p} = 2700k_{g} \text{ m}^{-3} \) is the mineral particle density, and \( \theta_{sat,min,i} \) is the porosity of the mineral soil.

The thermal conductivity of dry soil \( \lambda_{dry,i} \) (W m\(^{-1}\) K\(^{-1}\)) is calculated as follows:

\[
\lambda_{dry,i} = (1 - V_{soc})\lambda_{dry,min,i} + V_{soc}\lambda_{dry,om},
\]

where \( \lambda_{dry,min,i} \) is the thermal conductivity of dry mineral soil and is calculated as follows:

\[
\lambda_{dry,min,i} = 0.135\rho_{d,i} + 64.7 \frac{2.700 - 0.947\rho_{d,i}}{2.5 \times 10^{-3}}.
\]

where \( \rho_{d,i} = 2.700(1 - \theta_{sat,i})(1 - V_{g}) + 2.650\left(V_{g}\right) \) (Russo, 1983).

The Kersten number \( K_{e,i} \) (Kersten, 1949) is calculated as follows:

\[
K_{e,i} = \left\{ \begin{array}{ll}
\exp \left[ k_{f}\left(1 - \frac{V_{soc}}{S_{r,i}}\right)\right] & T_{r} \geq T_{f} \\
S_{r,i} & T_{r} < T_{f}
\end{array} \right. ,
\]

where \( S_{r,i} \) is the wetness of the soil with respect to saturation.

The volumetric heat capacity \( c_{i} \) (J m\(^{-3}\) K\(^{-1}\)) for soil (de Vries, 1963) is as follows:

\[
c_{i} = c_{s,i}\left(1 - \theta_{sat,i}\right) + \frac{w_{ic,i}}{\Delta z_{i}} C_{ic} + \frac{w_{liq,i}}{\Delta z_{i}} C_{liq},
\]

where \( C_{ic} \) and \( C_{liq} \) are the specific heat capacities (J kg\(^{-1}\) K\(^{-1}\)) of ice and liquid water, respectively. \( c_{s,i} \) is the heat capacity of soil solids, calculated as follows:

\[
c_{s,i} = (1 - V_{soc})c_{s,om,i} + V_{soc}c_{s,om},
\]

where \( c_{s,om} = 2.5 \times 10^{6} \text{ J m}^{-3} \text{ K}^{-1} \), and \( c_{s,om,i} \) is the heat capacity of mineral soil solids, calculated as follows:

\[
c_{s,om,i} = \frac{2.128\% \text{ sand}_{i} + 2.385\% \text{ clay}_{i} + 220\left(V_{g}\right)}{(\% \text{ sand}_{i} + \% \text{ clay}_{i} + \left(V_{g}\right)) \%} \times 10^{6}.
\]
3.2.2. Hydraulic Parameterization of Test3 Scheme

The porosity \( \theta_{sat,i} \) is as follows:

\[
\theta_{sat,i} = (1 - V_{soc}) \theta_{sat,min,i} + V_{soc} \theta_{sat,com},
\]

where the porosity of the mineral soil \( \theta_{sat,min,i} \) is as follows:

\[
\theta_{sat,min,i} = 0.489 - 0.00126(\%_{sand}) - 0.489(V_s)_i,
\]

(35)

The exponent \( B_i \) is as follows:

\[
B_i = (1 - V_{soc}) B_{min,i} + V_{soc} B_{com},
\]

where \( B_{com} = 2.7 \).

The exponent \( B \) of the mineral soil is as follows:

\[
B_{min,i} = 2.41 + 0.129(\%_{clay}) + 6.0(V_s)_i,
\]

(36)

The soil matric potential (mm) \( \Psi_i \) is as follows:

\[
\Psi_i = \Psi_{sat,i} \left( \frac{\theta_i}{\theta_{sat,i}} \right)^{-B_i},
\]

(37)

where the saturated soil matric potential (mm) \( \Psi_{sat,i} \) (Farouki, 1981; Letts et al., 2000) is as follows:

\[
\Psi_{sat,i} = (1 - V_{soc}) \Psi_{sat,min,i} + V_{soc} \Psi_{sat,com},
\]

(38)

where the saturated mineral soil matric potential \( \Psi_{sat,min,i} \) (Cousin et al., 2003; Pan et al., 2015) is as follows:

\[
\Psi_{sat,min,i} = -10.0 \times 10^{1.88 - 0.0131(\%_{sand})} \times (1 - V_s)_i + 130(V_s)_i,
\]

(39)

The saturated hydraulic conductivity for mineral soil \( k_{sat,min} \) (Cosby et al., 1984; Peck et al., 1979) is as follows:

\[
k_{sat,min} = 0.0070556 \times 10^{-0.884 + 0.153(\%_{sand})} \times \frac{2(1 - V_s)_i}{2 + (V_s)_i},
\]

(40)

The main modification equations of all the above schemes are shown in Table 2.

It is important to point out that due to the different influencing factors of hydrothermal parameters in different schemes, even if the measured soil organic carbon mass contents and soil gravel mass proportions is zero, the changes in hydrothermal parameters are quite different.

4. Soil Thermal Properties

4.1. The Changes in Soil Thermal Parameters

The soil thermal parameters modified by different parameterization schemes have changed considerably. Figure 3 shows the changes in soil thermal parameters in the test1, test2, and test3 schemes compared with the original CLM scheme at the QOMS station. Figures 4 and 5 show the thermal parameter changes at the BJ station and A’rou station, respectively. For the underlying surface of the Gobi (QOMS, Figure 3), the variations in soil temperature from BJ in situ observations (black), CLM (blue), test1 (orange), test2 (green), and test3 (red) at 5, 10, 20, and 40 cm depths from June to August 2019.
thermal conductivity of soil solids and porosity in test1 decreased with increasing soil depth, which made the saturated thermal conductivity of shallow soil decrease, and the deeper soil increase compared to the CLM scheme. The thermal conductivity of dry soil increased with increasing depth and led to an increase in the soil thermal conductivity. And the volumetric content of gravel in soils in test1 made the heat capacity of mineral soil solids and heat capacity of soil solids increase. The soil thermal conductivity and heat capacity of soil solids in test2 were reduced compared to those of the CLM scheme. The change in test3 was basically the same as that in test2, but the change range was different. At station BJ (Figure 4), the gravel content of the alpine meadow was approximately one-tenth of that of the Gobi (Table 1). Compared with the CLM scheme, the soil thermal conductivity and heat capacity of soil solids in test1 had a relatively small increase or decrease. However, the organic carbon content of the BJ station was 4 times that of the QOMS station, which greatly reduced the thermal conductivity of dry soil, heat capacity of soil solids, and soil thermal conductivity in test2. The change range of test3 for the thermal conductivity of dry soil was smaller than that of test2, which made the soil thermal conductivity decrease slightly. At station A’rou (Figure 5), test1 taking into account the influence of quartz and other minerals greatly reduced the mineral soil solid thermal conductivity. Compared with the CLM scheme, the gravel content of the surface layer was almost zero, so that the porosity was almost unchanged, leading to the saturated thermal conductivity reduced. Moreover, test2 and test3 (the organic carbon content was 2–3 times that of the alpine meadow) had a greater change than that of the alpine meadow. The increasing proportion of soil organic matter (socEV) and almost zero volumetric content of gravel (Vg) in soils made the heat capacity of mineral soil solids almost unchanged and the heat capacity of soil solids had a greater change.

By analyzing the changes in the thermal parameters of tests 1–3 on different underlying surfaces, it was found that the gravel scheme (test1) mainly increased the saturated thermal conductivity and the thermal conductivity of dry soil, increasing the soil thermal conductivity and heat capacity of soil solids. The organic carbon scheme (test2) mainly reduced the saturated thermal conductivity and the thermal conductivity of dry soil so that the soil thermal conductivity decreased. And the increasing proportion of soil organic matter made the heat capacity increased. The changes in the comprehensively considered test3 were basically the same as those in test2, and the change range was slightly different.

### 4.2. Soil Temperature

Figure 6 depicts the variations in soil temperature from in situ observations, CLM, test1, test2, and test3 at the QOMS station from June to August 2016. Figures 8 and 10 show the BJ station and A’rou station from June to August 2019. Overall, the variation in soil temperature at different depths was roughly the same, and the surface soil temperature fluctuated significantly. As the depth increased, the soil temperature tended to stabilize at 40 cm because the surface soil temperature is more sensitive to changes in solar radiation, wind speed, precipitation and other factors. In general, the simulation results described the amplitude of the variations well, but the fluctuation range was larger than the in situ observations. Figure 7 shows the diurnal variations in soil temperature at the QOMS station from June to August 2016. Figures 9 and 11 show the results for the BJ station and A’rou station from June to August 2019, respectively. The diurnal changes in soil temperature in each layer at the QOMS station (Figure 7) showed that the CLM simulation and test1 had a large deviation from the in situ observations, and compared with the daily variation in
observed values, CLM and test1 had a time lead of 2–3 hr test2 and test3 corrected the abovementioned time error to a certain extent and shortened the error to 1–2 hr test1 and test3 were closer to the observed values for soil temperature at the 40 cm depth. The soil temperature changes at the BJ station showed that the average deviation of the surface simulation soil temperature in the CLM scheme and test1 was approximately 7 °C relative to the observation, while the error of test2 and test3 relative to the observed value was reduced to 2.5 °C; other soil layer errors were reduced by more than half. The CLM and test1 schemes of the BJ station (Figure 9) also had the change time advance phenomenon, with a time error of approximately 1 hr. After the parameterization schemes of test2 and test3 were revised, test2 and test3 had a time lag of approximately 0.5 hr compared with the observed value. For station A’rou (Figure 11), the CLM and test1 differed greatly from the in situ observations at each layer, and there was still a phenomenon of a changed advance time. Test2 and test3 corrected the abovementioned time error to a certain extent, especially for the 20 cm depth soil, so that the time error was reduced by 2.5 hr, and the amplitude of daily variations was consistent with the observations. For the soil temperature at a depth of 40 cm, CLM and test1 had larger oscillations with respect to the in situ measurements, while the simulated values of test2 and test3 were closer to the observed values. Figure 12 shows the Taylor diagrams illustrating the simulated soil temperature at the 5 cm, 10 cm, 20 cm and 40 cm depths versus the in situ observations for CLM, test1, test2, and test3 at the QOMS (a), BJ (b), and A’rou (c) stations. At the QOMS station (Figure 12a), these results emphasize that the relative bias of the simulated values is less than 10% and that the four schemes have a good range of correlations at different soil levels. Most of the values are higher than 0.60. The standard deviation of the observed soil temperature at 10 cm depth is 2.7 °C, the standard deviations of the simulated values are all greater than 4 °C, the variability is greater than the observed value, and the amplitude of the variations is larger. As the depth increases, the standard deviation is closer to the observed value. The correlation coefficients are all greater than 0.9 at the BJ station (Figure 12b), passing the 5% significance level. The relative biases of test2 and test3 are smaller, and the standard deviation is closer to the observed value. For station A’rou (Figure 12c), the revision parameterization schemes at a depth of 5 cm increase the error between the simulated value and the observed value. Test2 and test3 performed better at other levels. In view of the characteristics of the E value, this paper directly used it as an indicator to evaluate the simulation effects of the four schemes. A smaller E value indicates a smaller standard deviation and a greater correlation with the observed value. Table 3 shows the E values of CLM, test1, test2 and test3 at different soil depths. In the aspect of simulated soil temperature of the shallow soil depth (0–20 cm), test2 and test3 considering organic carbon were better (except for the A’rou station at a 5 cm depth), while test1 and test3 considering the gravel were better in the deeper layer (40 cm). Moreover, the places where deeper gravel content was zero in test2 and test3 were better and closer to the observed value.

The reason for the differences between the different schemes lies in the different changes in the revised soil thermal parameters. The difference in the shallow soil between test1 and the observation was mainly due to the overestimation of the mineral soil solid thermal conductivity and thermal conductivity of dry mineral soil, which ultimately overestimated the soil thermal conductivity, resulting in larger fluctuations and errors. Test2 and test3 reduced the soil thermal conductivity and increased the heat capacity of soil solids so that the amplitudes of the variations in the simulations were reduced and closer to the observed

![Figure 10. Variations in soil temperature from A’rou in situ observations (black), CLM (blue), test1 (orange), test2 (green), and test3 (red) at 5, 10, 20, and 40 cm depths during June to August 2019.](image-url)
values. Except for the zero gravel content of station A’rou, the gravel content in the deep layer was larger, which caused the thermal conductivity of dry soil in test2 to be underestimated, resulting in a larger deviation; thus, the schemes of test1 and test3 were better. By comparing the main changes in soil thermal parameters, it was found that for the underlying surface of the Gobi and alpine meadows, the changes in soil thermal conductivity were more significant, and the revision of the thermal conductivity of dry soil was more critical. On grassland where the gravel content was almost zero, the changes in soil thermal parameters were almost the same, mainly due to the influence of soil moisture content on the soil thermal conductivity. Soil water is the carrier of heat, and increasing soil water content enhances the soil’s ability to absorb heat, resulting in increased soil temperature. Because of the lush surface vegetation at station A’rou and the relatively larger shallow soil water content, the simulations underestimated the shallow soil water content, thereby underestimating the soil thermal conductivity and increasing the error of the revised results.

5. Soil Hydraulic Properties

5.1. Changes in Soil Hydrological Parameters

Figure 13 shows the changes in soil hydrological parameters in the test1, test2, and test3 schemes compared with the original CLM scheme at the QOMS station. Figures 14 and 15 show the hydrological parameter changes of the BJ station and A’rou station, respectively. For the QOMS station (Figure 13), the decreasing porosity and increasing exponent $B$ caused the saturated soil matric potential and saturated hydraulic conductivity to decrease in test1 and test3, and the amplitude of variations increased with increasing soil depth. In contrast, the porosity decreased and the exponent $B$ increased in test2 compared with the original CLM scheme, which caused the saturated soil matric potential to increase and the saturated hydraulic conductivity to decrease. At station BJ (Figure 14), the reduced porosity and exponent $B$ decreased the saturated soil matric potential and saturated hydraulic conductivity. Due to the significant decrease in exponent $B$ in test1, the amplitude of decreasing variations in the saturated soil matric potential and saturated hydraulic conductivity was much smaller than those at the QOMS station. In terms of the test2 and test3 schemes, the porosity increased and exponent $B$ decreased, which reduced the saturated soil matric potential and increased the saturated hydraulic conductivity. The change trend of test2 and test3 at station A’rou (Figure 15) was the same as that at station BJ, but the amplitude of variation was slightly different, while the main change of test1 was the sharp decrease of exponent $B$.

By analyzing the changes in the hydraulic parameters on different underlying surfaces, the gravel scheme (test1) mainly affected the changes in soil water content by reducing porosity and increasing exponent $B$ to reduce the saturated soil matric potential and saturated hydraulic conductivity. The organic carbon scheme (test2) mainly increased porosity and decreased exponent $B$ to increase the saturated hydraulic conductivity and change the soil water content. Test3 mainly increased porosity and decreased exponent $B$, which increased the saturated hydraulic conductivity, and the change in the saturated soil matric potential was closely related to the soil texture.
Figure 12. Taylor diagrams illustrating the simulated soil temperature at 5 cm (black), 10 cm (red), 20 cm (blue), and 40 cm (orange) depth versus in situ observations for CLM (number 1), test1 (number 2), test2 (number 3), and test3 (number 4) at stations QOMS (a), BJ (b), and A'rou (c). O-5/10/20/40 represent the standard deviation of the observed soil temperature at soil depths of 5, 10, 20, and 40 cm, respectively.

Table 3

| $E$ (cm) | QOMS  | BJ   | A'rou |
|----------|-------|------|-------|
|          | CLM   | test1 | test2 | test3 | CLM   | test1 | test2 | test3 | CLM   | test1 | test2 | test3 |
| 5        | 3.73  | 3.59  | 2.99  | 3.00  | 2.76  | 2.93  | 1.50  | 1.48  | 3.81  | 4.36  | 4.17  | 4.32  |
| 10       | 3.72  | 3.59  | 2.99  | 3.00  | 2.76  | 2.93  | 1.50  | 1.48  | 3.81  | 4.36  | 4.17  | 4.32  |
| 20       | 2.52  | 2.35  | 1.67  | 1.65  | 1.90  | 1.76  | 1.18  | 1.19  | 2.13  | 2.17  | 1.17  | 1.16  |
| 40       | 1.50  | 1.08  | 1.45  | 1.23  | 2.16  | 1.53  | 1.67  | 1.65  | 0.63  | 0.58  | 0.45  | 0.44  |
5.2. Soil Moisture

Figure 16 depicts the variations in soil moisture from in situ observations, CLM, test1, test2, and test3 at the QOMS station from June–August 2016. Figures 17 and 18 show the results for the BJ station and A’rou station from June to August 2019, respectively. From the perspective of the changes in the simulation results, different parameterization schemes could well capture the small changes in the observed soil moisture. The surface soil moisture was affected by less precipitation and evaporation was as low as 0.06 m$^3$ m$^{-3}$ at the QOMS station (Figure 16), while the soil moisture at the 40 cm soil depth was as low as approximately 0.025 m$^3$ m$^{-3}$ due to poor water retention performance with increasing gravel content. The maximum soil moisture was approximately 0.13 m$^3$ m$^{-3}$ at the 20 cm depth. In general, the original CLM scheme data were relatively higher than the simulated soil moisture data, and tests 1–3 improved the simulation accuracy at deeper soil levels. Figures 16d–16f shows that the daily variation in soil moisture in each layer remained almost stable with no fluctuations, and the improved results of the four schemes could be seen.
more intuitively, especially compared with the other schemes. The simulation-observation error of test3 at the 20–40 cm depth was shortened by more than half. With respect to the BJ station variations (Figure 17), the soil moisture simulation at 0–10 cm fluctuated greatly, which was lower than the observations, while the soil moisture at the 20–40 cm depth simulation was larger than the observed value. The soil moisture at the 20 cm depth was the smallest, at approximately 0.095 m$^3$m$^{-3}$. Compared with the other soil layers, the error between the simulation results and the observations was the smallest. Due to the vigorous growth of vegetation and high water-holding capacity of soil at station A’rou (Figure 18), the average surface soil moisture was larger than that of the other two stations, which was approximately 0.3 m$^3$m$^{-3}$. Overall, the four simulation scheme data were lower than the observed values, especially the test1 scheme, which was approximately 0.15 m$^3$m$^{-3}$ lower than the observed values. The diurnal variation showed the difference between the original CLM and test2 schemes, the observed values decreased with increasing soil depth, and other scheme errors were also reduced to varying degrees. From Figures 16–18, we can see that the surface soil moisture has a sharp variation. This may be due to the increase in precipitation that caused the surface

Figure 14. The same as Figure 13, but for station BJ.
soil moisture to rise rapidly, but when the precipitation stopped and the wind speed was high, the surface soil evaporation increased, causing the soil moisture to decrease (Figure 2).

In addition, the simulation results showed that the improvement of soil hydraulic parameters had a positive feedback effect on the soil temperature, while the improvement of thermal parameters had little effect on the soil moisture. This result was because soil water is the carrier of soil heat. Soil temperature can quickly respond to water changes, the latent heat of phase change of soil water affects temperature changes, and soil water content is mainly determined by precipitation, run-off and groundwater. Therefore, it is very important to design an appropriate parameterization scheme to balance the relationship between the thermal and hydraulic parameters.

Figure 19 shows the Taylor diagrams illustrating the simulated soil moisture at depths of 5, 10, 20, and 40 cm versus in situ observations for CLM, test1, test2, and test3 at stations QOMS (a), BJ (b), and A’rou (c). The correlation coefficient of the shallow simulated soil moisture at the QOMS station (Figure 19a) was greater than 0.9, while those of the deep soil were greater than 0.8, which passed the 5% significance level and had a
good correlation. The simulated standard deviation at the 20 cm layer was the closest to the O-20 line, and the amplitude of the variations was basically the same as in the in situ observations. Compared with observations, the depths of 10 and 40 cm had higher variability and larger deviations, especially for the simulation results at the 40 cm depth, and test3 and test2 greatly reduced the error and performed well. With respect to station BJ (Figure 19b), test2 and test3, with smaller negative relative biases at the 0–10 cm depth, performed better, while test1 and test3, which had smaller positive relative biases and standard deviations similar to the observed value at the 20–40 cm depth, performed better and were consistent with the change in the observed value. The relative biases at station A’rou (Figure 19c) were all less than 0, indicating that the simulations were somewhat lower than the observed values. The standard deviations of test1 and test3 at the 10 and 20 cm depths were closer to the observed values, and the variability was smaller. The four parameterization schemes had good correlations at different soil levels, and the correlation coefficients were all higher than 0.7. Combined with Table 4, it can be concluded that the test1 and test3 schemes that considered the influence of gravel performed better for the underlying surface of the Gobi and alpine meadows at the 20–40 cm depths and were closer to the observations. For the grassland underlying the surface, the original CLM scheme and test2 performed better. The results at the 0–10 cm depth were more affected by external forcing and soil texture, and further discussion is needed.

The soil hydraulic parameter changes led to the revised difference of parameterization schemes. For soils at the 20–40 cm depth, test1 and test3 reduced porosity, saturated soil matric potential, and saturated hydraulic...
The soil hydraulic conductivity due to the influence of gravel, making the soil moisture lower and causing the model to perform better. The soil hydraulic conductivity is a function of the saturated hydraulic conductivity, porosity, soil water content and exponent $B$. Therefore, the changes of the Gobi in soil hydraulic conductivity of test1 and test3 were almost the same, and the modified soil matric potential became a key parameter. Moreover, the correction of exponent $B$ contributed greatly to the revision of soil matric potential. On the underlying surface of the alpine meadow, the test1 scheme greatly underestimated the soil hydraulic conductivity, which was one order of magnitude different from that of test3; thus, the soil hydraulic conductivity should be the focus of consideration for areas with high organic carbon content. In grasslands, increasing porosity and soil hydraulic conductivity enhanced the water-holding capacity, resulting in higher soil moisture and compensating for the shortcomings of the overall smaller simulation value. As a consequence, the change in porosity had a greater impact on the correction of soil thermal conductivity in test2, which in turn affected the change in soil moisture content. Due to the large deviations in the surface simulation results of Gobi
and alpine meadows, it is still necessary to continue to discuss their revised parameterization schemes in future research.

6. Summary and Discussion

In this study, the QOMS, BJ, and A’rou observation stations, representing different typical underlying surfaces of the TP, were used to study the soil thermal properties, soil hydraulic properties, soil temperature, and soil moisture in the plateau area. As an important part of the soil, gravel has an important impact on soil thermal and hydraulic properties; the plateau area has poor soil development and a thin soil layer, but the soil organic carbon content in the southeast area is relatively high, and the soil texture is relatively coarse. Therefore, three parameterization schemes, one considering gravel influence (test1), one considering organic carbon influence (test2), and one considering the comprehensive effect (test3), were set up to modify different typical underlying surfaces of the TP. In addition, their soil thermal properties and hydraulic property variations were discussed. This research discussed the key thermal and hydraulic parameters that affected the performances of different schemes from the perspective of observation data to improve the simulation ability of soil temperature and soil moisture in plateau areas. The main results are as follows:

1. In terms of soil thermal properties, test1 mainly increased the saturated thermal conductivity and the thermal conductivity of dry soil so that the soil thermal conductivity and heat capacity of soil soils increased, which ultimately led to an increase in soil temperature. The revision of hydraulic properties

Figure 18. The same as Figure 16 but for station A’rou from June to August 2019.
Figure 19. Taylor diagrams illustrating the simulated soil moisture at 5 cm (black), 10 cm (red), 20 cm (blue) and 40 cm (orange) depths versus in situ observations for CLM (number 1), test1 (number 2), test2 (number 3), and test3 (number 4) at stations QOMS (a), BJ (b), and A’rou (c).

Table 4

| E (cm) | QOMS | BJ | A’rou |
|--------|------|----|-------|
|        | CLM  | test1 | test2 | test3 | CLM  | test1 | test2 | test3 | CLM  | test1 | test2 | test3 |
| 5      | –     | –     | –     | –     | 0.042 | 0.043 | 0.039 | 0.039 | 0.044 | 0.048 | 0.046 | 0.049 |
| 10     | 0.048 | 0.059 | 0.050 | 0.053 | 0.030 | 0.029 | 0.029 | 0.028 | 0.029 | 0.033 | 0.030 | 0.034 |
| 20     | 0.046 | 0.039 | 0.043 | 0.043 | 0.033 | 0.031 | 0.033 | 0.029 | 0.027 | 0.028 | 0.027 | 0.028 |
| 40     | 0.115 | 0.052 | 0.113 | 0.039 | 0.040 | 0.039 | 0.045 | 0.040 | –     | –     | –     | –     |
mainly occurred by reducing porosity and increasing exponent $B$ to reduce the saturated soil matric potential and saturated hydraulic conductivity, thereby reducing soil moisture.

2. The thermal properties in test2 were mainly reduced by reducing the saturated thermal conductivity and the thermal conductivity of dry soil so that the soil thermal conductivity was reduced and the heat capacity was increased, resulting in a lower soil temperature simulation and smaller amplitude of variation. For the revision of hydraulic properties, the saturated hydraulic conductivity was increased by increasing porosity and decreasing exponent $B$, thereby increasing soil moisture.

3. The test3 scheme was the same as test2 for the modified changes in thermal properties, except that the proportion of change was slightly different. The revision of hydraulic properties was mainly to increase the saturated hydraulic conductivity by increasing the porosity and reducing exponent $B$. The change in the saturated soil matric potential was closely related to the soil texture.

4. Regarding the soil temperature simulation, test2 and test3 were better for the shallow soil (0–20 cm), while test1 and test3 were better for the deeper (40 cm) soil simulation. In the area with almost zero gravel content, test2 and test3 performed better and were closer to the in situ observations. For the underlying surface of the Gobi and alpine meadows, the soil thermal conductivity changed more significantly, and the correction of the thermal conductivity of dry soil was more critical. The change in soil water content and its influence on the soil thermal conductivity were more critical on the underlying surface of the grassland.

5. In terms of soil moisture simulation, the test1 and test3 schemes performed better on the Gobi and alpine meadows at the 20–40 cm depth. As the changes in soil hydraulic conductivity of test1 and test3 were almost the same as those of Gobi, the modification of soil matric potential became a key parameter, and the variations in exponent $B$ contributed greatly to the correction of the soil matric potential. In addition, soil hydraulic conductivity should be the focus of consideration for alpine meadows. For the grassland underlying surface, the original CLM scheme and test2 performed better. The change in porosity had a greater impact on the correction of soil thermal conductivity, which in turn affected the change in soil moisture content.

Although the new parameterization scheme improved the simulation effect of different underlying surface observation stations to different degrees, there were still some problems: (a) For the simulation of soil moisture on different underlying surfaces of the plateau at the 0–10 cm depth, there was still a large deviation from the observed value, which may have been related to the improved surface soil moisture equation of CLM5.0. (b) The overall applicability of the improved parameterization schemes on the TP needs to be verified. Improving the universality of parameterization schemes and making the simulation results closer to the real value are our goals in future research.

**Data Availability Statement**

Datasets for this research are included in this paper: C. L. Ma, Lü, and Pan (2020); Y. Ma, Hu et al. (2020); and Liu et al. (2020). (https://data.tpdc.ac.cn/zh-hans/data).

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