Investigation on the soil thermal conductivity of different land surface patterns in the northern Qinghai-Tibetan Plateau, China

Ren Li, Tonghua Wu, Lin Zhao, Changwei Xie, Yao Xiao, Guojie Hu & Yizhen Du

Cold and Arid Regions Environmental and Engineering Research Institute, Chinese Academy of Science, Lanzhou, Gansu, China

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ABSTRACT
Several numerical methods are applied and compared in the calculation of soil thermal conductivity (STC) in the Kekexili (QT01), Beiluhe (QT02), Kaixinling (QT05), and Tongtianhe (QT06) regions in the northern Qinghai-Tibetan Plateau based on data from January 2004 to December 2013. The results show that the STC of the active layer in the study region exhibited marked seasonal variations: it was low during the cold season but high in the warm season. Averagely, the mean value was 1.080 Wm^{-1}K^{-1}, ranging from 0.752 to 1.371 Wm^{-1}K^{-1}. In the frozen state (FS), STC was 0.955 W.m^{-2}.K^{-1} while in the unfrozen state (UFS), it was 1.204 W.m^{-2}.K^{-1}. STC increased with increasing soil bulk density but decreased with increasing vegetation cover of the land surface; STC of alpine frost meadow soil was greater than that of alpine frost steppe soil; fine-grained soil with low unfrozen water content and a low saturation degree resulted in low STC in the cold season; and monthly mean STC can be well expressed as a function of conventional meteorological data. Verification results further ensured that the proposed model accurately predicts monthly STC values.

RÉSUMÉ
Plusieurs méthodes numériques sont appliquées et comparées dans le calcul de la conductivité thermique des sols (STC) dans les régions Kekexili (QT01), Beiluhe (QT02), Kaixinling (QT05), et Tongtianhe (QT06) dans le nord du plateau Qinghai- Tibet en se basant sur les données de janvier 2004 à décembre 2013. Les résultats montrent que la STC de la couche active dans la région étudiée a exposé des variations saisonnières marquées : c'était bas pendant la saison froide, mais haut durant la saison chaude. En moyenne, la valeur était de1.080 Wm-1K-1, pour un intervalle de données allant de 0.752 à 1.371 Wm-1K-1. Pour l'état congelé, la STC était de 0.955 W.m-2.K-1, alors que pour l'état non-gelé elle était de 1.204 W.m-2.K-1. La STC augmente avec l'augmentation de la masse volumique apparente du sol, mais diminue avec l'augmentation de la couverture de végétation à la surface du sol; La STC d'un sol de prairie alpine gelée était plus élevée que celui de la steppe alpine gelée; les sols à grains fins ayant un faible contenu en eau non-gelée et un faible degré de saturation ont une STC basse dans la saison froide; et la moyenne mensuelle des STC peut être exprimée comme une fonction de données météorologiques conventionnelles. Les résultats de la vérification ont assuré que le modèle proposé prédit précisément les valeurs de STC mensuelles.

1 INTRODUCTION

Soil thermal conductivity (STC) (W.m^{-1}.K^{-1}) is an important thermal property relating to the ability of heat exchange at the ground surface or of the penetration, amplitude and effects of the daily and seasonal temperature variations (Farouki 1981). It is one of the essential parameters for determining soil freezing and thawing depths and engineering stability (Farouki 1981; Inaba 1983; Tarnawski and Wagner 1993). During the freezing-thawing processes of the active layer, the STC controls the response of energy and moisture exchanges between the ground surface and permafrost (Overduin et al. 2006). It has been shown to have a significant impact on the partitioning of surface energy fluxes and the prediction of soil and skin temperatures (Peters-Lidard et al. 1998). It is also a crucial parameter for the land surface model (Ye et al., 1991; Zhang et al., 2003; Yang et al. 2005) in determining the surface energy and water budget. Relatively precise STC of the active layer in permafrost regions are important data for predicting global changes and a prerequisite for the design and construction of engineering constructions in cold regions (Ling and Zhang 2004; Xin et al. 2012; Xiao et al. 2013).

STC and the prediction of it has become of interest as a widely-used parameter in land surface processes. It is well known that STC is a natural property of soil; its value is affected by soil composition, structure and temperature. Among the many factors that influence STC (Farouki 1981), such as mineral composition, bulk density, soil porosity, and soil moisture content (SMC) (m^{3}m^{-3}), SMC is the one that varies most under field conditions. In situ monitoring of STC as a function of SMC can represent a significant challenge. Consequently, much effort has been made to develop STC models based on easily measurable factors. However, it still needs to be improved to cover a wider range of soil types, porosity, and degrees of saturations. All of these above-mentioned results were based on the laboratory condition. Tarnawaki and Wagner (1993) pointed out that the STC measured in the laboratory with disturbed samples could not reflect those in the natural environment. Therefore, numerous measurements must be conducted to properly characterize a given site. As for the permafrost region, the condition is more complicated due to the fact that the phase of soil moisture is different under the frozen or thawed state. It is well knoen that STC of ice is three to four times that of water. In different stages of the freezing and thawing processes, the ratio of soil ice content and unfrozen water content in the active layer is continuously changing and the coefficient of heat conduction exhibits significant differences. Therefore, the active layer is viewed as a variable thermal resistance of the processes of exchange between permafrost and the atmosphere (Lachenbruch, 1994). Accordingly, the water-heat regime of the active layer in the permafrost region of the Qinghai-Tibetan Plateau (QTP) is of crucial concern to scholars and has been the subject of significant research (Ding et al., 2000; Zhao et al., 2000; Wu and Zhang, 2010; Wu et al., 2010; Zhao et al., 2010; Wang et al., 2005; Li et al., 2005; Li et al., 2010; 2012; Luo et al., 2009).

Existing research results and accumulated data provide a solid basis for profound understanding of the water-heat transfer mechanism in the freezing and thawing processes of the active layer. However, those studies have been confined to only the thermal properties of the land surface pattern; little work has dealt with the comparative study of the thermal properties of different land surfaces pattern in the freezing and thawing environment of the active layer. The purpose of this study is to recognize the in situ STC in the QTP. In this study,the observational data from an overall freezing-thawing cyclic period of the active layer in the permafrost region of the northern QTP (2004–2013), combined with the vegetation conditions of the various land surface patterns, are used to determine STC of topsoil the active layer associated with different land surface pattern. After that, an statistical model to estimate STC is provided using the routine meteorological data. All these will provide the corresponding land surface parameters for numerical simulations of the freezing-thawing processes of the active layer.

2 DATA AND METHODS

Four study regions, Kekexili (QT01), Beiluhe (QT02), Kaixinling (QT05), and Tongtianhe (QT06) in the northern QTP were selected. These regions are located in the continuous permafrost zone on the high plain in the source area of the Changjiang River, with a semiarid climate of the Plateau's subfrigid zone (Lin and Wu, 1981). The mean altitude is 4,673 m and the area comprises alpine frost steppe and alpine frost meadow (Zheng et al., 1979). The vegetation cover at the four observation sites varies between 20% and 90% and the dominant plant species is Carex moorcroftii. The shallow surface layer (0–10 cm) is predominantly silty soil. The annual mean air temperature in the study regions is –5.5 ~ 3.1 °C and the annual mean relative humidity is 55.2 ~ 53.3%. According to the multiyear climatic data recorded at the Wudaoliang and Tuotuohe meteorological stations, the annual precipitation in the region is approximately 330 mm and mainly falls in the summer monsoon period (90% of the annual total precipitation falls between May and September); the maximum cumulative precipitation occurs between June and August. The permafrost thickness in the QT01 and QT05 regions is 80–100 m, and in the QT02 and QT06 regions it is 40–60 m (Wang et al., 1979). The active layer thickness at the QT01, QT02, QT05, and QT06 observation fields are 1.63 m, 2.42 m, 2.88 m, and 2.52 m, respectively (Zhao et al., 2010). The geographical locations and land surface conditions of the various stations are presented in Table 1.

Table 1 Geographical locations of the study region

No.	Station name	Longitude (E)	Latitude (N)	Altitude (m)	Underlying surface condition	Vegetation cover (%)
QT01	Kekexili	93°03′	35°09′	4,734	Alpine frost meadow	80–90
QT02	Beiluhe	92°55′	34°49′	4,656	Swamp meadow	60–80
QT05	Kaixinling	92°22′	33°57′	4,652	Desert steppe	20–30
QT06	Tongtianhe	92°14′	33°46′	4,650	Alpine frost steppe	40–50

The utilized data include soil heat flux, soil temperature, and moisture gradient observations at the 0–10-cm soil layer in the QT01, QT02, QT05, and QT06 observation fields during the period from October 2003 to September 2004. Soil heat fluxes were measured using a HFPO1SC self-calibrating heat flux plate (Hukseflux USA, Manorville, New York), with an accuracy of ±3%. Soil temperatures were measured using a 105T thermocouple temperature sensor (Campbell Scientific, Inc., Ogden, Utah). Soil moisture contents were measured using a Hydra soil moisture sensor (Stevens Vitel Inc., Portland, Oregon). Surface temperature sensors were placed in the active layer at 2-, 5-, and 10-cm depths, and soil moisture sensors and heat flux plates were placed at 5- and 10-cm depths. All the instruments were attached to a CR23X data logger (Campbell Scientific, Inc., Ogden, Utah) and the data were collected once every hour.

Soil thermal conductivity (STC) can be calculated by the following formula (Wang et al., 2005; Li et al., 2010):

$$STC = -\frac{(G_1 + G_2)/2}{\delta T_s/\delta z} \qquad [1]$$

Where G1 and G2 are soil heat fluxes at 5-cm and 10-cm depths, respectively; δz is the thickness between the two heat flow plates (i.e., 5 cm); and $\delta Ts/\delta z$ is the temperature gradient between the 5-cm and 10-cm depths.

In comparing the various calculation schemes of STC, Farouki (1981) concluded that the Johansen scheme can better estimate soil heat conductivity. Studies by Peters-Lidard et al. (1998) also showed that of the various SVAT (soil-vegetation-atmosphere transfer) schemes, Johanson's method can better estimate soil heat conductivity. In this study we used the Johanson scheme to test and verify our calculations. On this basis, during the soil freezing period the results obtained in this study had a calculation error 7.8%; during the soil thawing period our calculation error was 9.6%.

For additional verification, on March 30, 2006 we used a KD2 portable soil heat detector (Decagon Devices, Inc., Pullman, Washington) to measure STC at the QT01 observation field. The active layer monitoring data of that same period were applied in Equation [1], and the obtained result had an error of 5.8% relative to the KD2 measured result. These test results show that the calculation scheme used in this study had relatively small calculation errors and could thus be used to estimate STC in the study regions.

Because no 0-cm surface temperature, T_0, was recorded at the corresponding study stations, here the 0-cm surface temperature was estimated using the scheme introduced in Zhao et al. (2008):

$$T_0 = \frac{5T_{s2} - 2T_{s5}}{3} \qquad [2]$$

Where Ts_2 and Ts_5 are the soil temperature at the 2-cm and 5-cm depths, respectively. We obtained the soil temperature data at 0-, 5-, 10-, 15-, and 20-cm depths recorded at the Wudaoliang meteorological station during the whole year of 2005, and we used the Lagrange interpolation method (Weng et al., 1983) to test and verify the above formula. The calculated results of the 0-cm surface temperature showed that at a day scale the absolute error of the calculated results using above scheme was 0.2 °C and the relative error was 6%. It can thus be seen that the use of the above scheme to estimate daily mean surface temperature is feasible.

The degree-day factors were calculated by the following method. That is, during a whole freeze-thaw cyclic period, when the daily mean temperatures at 5-cm depths were consistently lower than 0^{o}C, the frozen degree-day (FDDs) was calculated by accumulating the daily ground temperature. The thawed degree-day (TDDs) was calculated by the same way as the daily mean temperatures at 5-cm depths were consistently larger than 0^{o}C.

3 RESULTS AND DISCUSSION

3.1 Soil temperature changes of the active layer

Soil temperature is an important thermal index because its values reflect the soil heat changes. Our observed results show that the variations of soil temperature (Ts5) at the 5-cm depth in the QT01, QT02, QT05, and QT06 observation fields were similar: the Ts5 low value occurred in January and its high value occurred in July; from January to July the Ts5 value rose and from August onwards the Ts5 value gradually dropped. These variations of Ts5 at the four stations resulted from the combined actions of solar energy received by the land surface and the underlying surface.

The concrete data of Ts5, air temperature Ta, and soil degree-day factor (DDs) at the 5-cm depth during the freezing-thawing period at the four stations are presented in Table 2. The daily mean Ts5 < 0 °C, DDs is expressed as FDDs (°C.d) (Vermette and Christopher, 2008). It can be seen from Table 2 that the annual amplitudes of Ts5 and Ta at the QT05 site were 23.2 °C and 22.0 °C, respectively. The large annual amplitudes of Ts5 and Ta at the QT05 site were related to the sparse vegetation cover of the underlying surface. At the other three stations the annual amplitude of Ts5 was less than 20.0 °C and that of Ta was less than 21.4 °C. Comparison of Ts5 and Ta values at the four stations indicates that the temperature at the southern sites was higher than that of the northern sites. The degree-day factors at the four stations show that the TDDs varied between –457.3 ~ 780.1 °C·d during the thawing period of the active layer, while during the freezing period the FDDs varied between –2,179 ~ 1,849 °C·d. TDDs and FDDs values increased with decreasing latitude, and with the decrease in latitude the absolute value of the ratio of the two increased from 0.210 to 0.422.

Table 2 Temperature and degree-day factors at the four study stations

Station	T_{s5} (°C)			T_a (°C)			Degree-day (°C·d)		
	January	July	Year	January	July	Year	Amplitude	FDD_s	TDD_s
QT01	–12.2	7.6	–1.6	–15.6	5.0	–5.1	20.7	–2,179.6	457.3
QT02	–10.9	8.4	–0.8	–15.3	5.3	–4.5	20.6	–2,114.4	539.5
QT05	–11.4	11.8	0.8	–15.1	6.9	–3.5	22.0	–1,986.9	751.8
QT06	–8.9	8.8	0.3	–14.2	7.1	–3.2	21.3	–1,849.4	780.1

3.2 Soil thermal conductivity

Equation [1] was used to calculate the topsoil thermal parameters of the active layer at the four stations, including Beiluhe:

Figure 1 shows the monthly mean STC of the shallow active layer (5–10 cm) at the QT01, QT02, QT05, and QT06 observation fields over the time period from 2004 to 2013. It can be seen from the figure that the soil thermal conductivities of the shallow active layer at these four stations exhibited obvious seasonal variations; they were large in summer and autumn but small in winter. As for QT01, the thermal conductivity in the shallow active layer was low in the cold season (October–the following April) but high in the warm season (May–August), and the thermal conductivity in the cold and warm seasons differed by 0.146 $W.m^{-1}.K^{-1}$. Over the period considered, the mean STC was 1.016 $W.m^{-1}.K^{-1}$. Under the frozen state (FS), STC was 0.941 $W.m^{-1}.K^{-1}$ while under the unfrozen state (UFS) the STC was 1.120 $W.m^{-1}.K^{-1}$; the difference between the two was only 0.179 W.m-1.K-1.

The variation behaviours of STC in the other three observation fields were similar to those of QT01. STCs started to increase in January or February and reached their maximum values in October or November and then started to decrease. The increase in thermal conductivity at the three stations was related to the increase of soil moisture content in the shallow active layer. We know that the precipitation in the Plateau region is affected by the Plateau's monsoon; in our specific study region, 90% of the annual precipitation falls in May–September, and during this period the soil moisture content in the active layer correspondingly increases and thermal conductivity also increases.

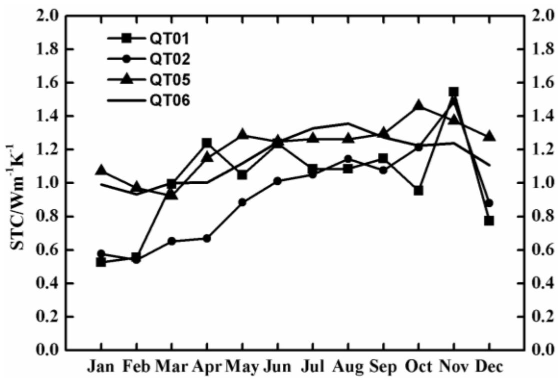

Figure 1 Seasonal variations of soil thermal conductivity over the time period from 2004 to 2013

STC is determined by the contact modes of various particles, soil properties, and soil moisture content (Weng et al., 1983). STC increases with increasing soil bulk density (Xu et al., 2001), but decreases with increasing soil organic matter content. From north to south in the study region, STC increased from 1.016 $W.m^{-1}.K^{-1}$ to 1.215 $W.m^{-1}.K^{-1}$ with decreasing latitude. Our analysis of soil bulk density in different observation fields (Table 3) indicates that from north to south the soil bulk density increases but soil porosity decreases, thereby increasing the compactness and close contact of soil particles. This is one of the important reasons for the difference in STCs in the different regions. Comparison of Table 1 and Table 2 shows that land surface types and vegetation coverage have a significant influence on STC (i.e., STC is lower in meadow areas than in steppe areas; similarly, STC was lower in the study area that had a heavy vegetation cover than in the study area having sparse vegetation cover). As demonstrates that the land surface types and vegetation conditions in different areas are also factors affecting thermal conductivity in arid and semiarid regions in QTP.

Table 3 Soil thermal conductivity, biomass, bulk density, and porosity at the study sites

Site	Thermal conductivity ($W.m^{-1}.K^{-1}$)			Biomass	Bulk density	Porosity
	Winter	Summer	Annual	(g/m^2)	(g/m^3)	
QT01	0.618	1.135	1.016	3287	1.250	0.528
QT02	0.666	1.069	1.028	2550	1.316	0.503
QT05	1.105	1.258	1.215	1115	1.340	0.494
QT06	1.011	1.309	1.152	2444	1. 416	0.466

The thermal conductivity of ice is almost four times that of water. Therefore, in general, STC in FS is high while that in UFS is low. As for the four stations in the study region, STC in FS was low, the same as in the adjacent Wudaoliang region (Li et al., 2005) of QT01 and the arid Gaize region (Wang et al., 2005) in the western QTP. The existing experimental study showed that in the –10 ~ 0 °C temperature range the accurate measurement of STC strongly depends on the soil temperature, the initial water content, and the particle size (Inaba,1983). In fine-grained soil, STC decreases as negative temperature drops. The unfrozen water in fine-grained frozen soils plays an import ant role in promoting heat transmission (Farouki,1981). As for the regions in this study located in a semiarid zone, precipitation there is scarce and the soil moisture content is relatively low (the land surface is sandy soil and silty soil, with poor water retention capability). Observation data during the study period show that at the beginning of freezing, the initial water content in the corresponding soil layer was 0.22 m^3m^{-3} in the QT02 observation field, while the initial water contents in the other three observation fields were less than 0.20 m^3m^{-3}. During the frozen period, the unfrozen water content in topsoil of the active layer in the study region was less than 0.08 m^3m^{-3}.As pointed by Farouki (1981) that the massive drop in STC is caused by the decreased unfrozen water content, and unfrozen water in fine-grained frozen soil may have a higher thermal conductivity than ice. The experimental results by Xu et al. (2001) confirmed that when soil moisture content is less than 0.20 m^3 m^{-3}, STC in UFS is greater than that of FS. During the soil freezing period in our study region, both the initial water content and unfrozen water content were small; therefore, the soils were only slightly saturated (with unfrozen water content less than 20%). The saturation degree is less than 30%. Existing studies show that when the actual saturation degree of fine-grained soil is less than the threshold value, the thermal conductivity decreases with a drop in temperature (Farouki, 1981; Lange and McKim, 1963; Haynes et al., 1980). Thus, fine-grained surface soil and low unfrozen water content in top soil in the active layer may be one of the important reasons for the low STC in FS in our study region in the QTP. In addition, frost heaving and ice lenses (Inaba, 1983) destroy soil structure, and thus the contact modes among the soil particles are changed. This is another possible cause of the reduced effective STC of frozen soil.

3.3 Effects of soil moisture content and other factors on thermal conductivity

Because soil is a medium consisting of three phases (solid, liquid, and gas), heat transfer occurs mainly by passing through soil particles, water, and air in arid and semiarid soils. Therefore, the magnitude of STC depends on the contact modes of various particles, the soil properties, and the soil moisture content (Weng et al., 1983). For a given soil, its mineral composition variation is relatively small and therefore has little effect on STC. Normally, the thermal conductivity of water is 0.57 $W.m^{-1}.K^{-1}$, whereas that of air is only 0.025 $W.m^{-1}.K^{-1}$; therefore, STC mainly depends on soil moisture content, and changes in the soil moisture content will alter the soil thermal parameters.

Figure 2 shows the mean thermal conductivity in relation to moisture content changes in an overall freezing-thawing cyclic period in the QT01, QT02, QT05, and QT06 observation fields. From the curve variations it can be seen that the soil moisture content has a significant effect on thermal conductivity. In winter, as the surface layer of the active layer starts to freeze from top to bottom, soil water moves toward the freezing front (Zhao et al., 2008). In the soil freezing processes, the unfrozen water in soil is a function of soil temperature: as soil temperature drops, the unfrozen water content in the shallow soil layer decreases exponentially (Xu et al., 2001). As described above, in unsaturated fine-grained soils the unfrozen water content in the soil has a significant effect on thermal conductivity, and the thermal conductivity of the surface layer of the active layer decreases rapidly with decreasing unfrozen water content in the surface soil layer. In the warm season, as the land surface temperature gradually rises, topsoil starts to thaw, soil moisture gradually increases, and thermal conductivity also increases. As the Plateau's summer monsoon sets in, precipitation increases and soil moisture sharply increases, and thermal conductivity also increases. The variations of these parameters can be described by the linear relation:

$$STC = 0.798 + 1.3904\theta_u \qquad [3]$$

Where θ_u is the soil volumetric water content (in winter it is unfrozen water content). The correlation coefficient of the two is 0.94, and the standard deviation is 0.064 W.m-1.K-1 at the 0.001 significance level. This shows that the correlation between STC and soil moisture content is statistically significant.

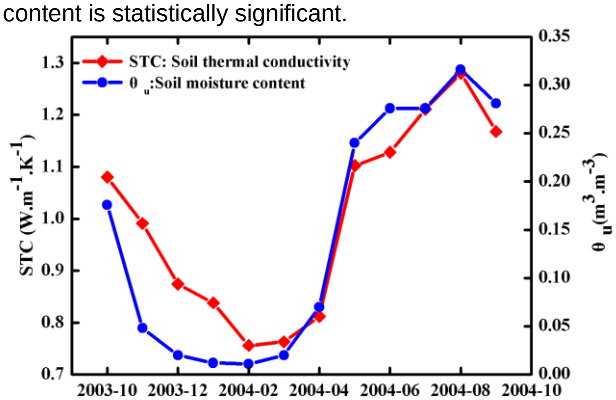

Figure 2 Effect of soil moisture content on thermal conductivity

STC and the temperature gradient determine the magnitude of soil heat flux, and in the land surface heat balance equation, changes of the soil heat flux value affect the land surface sensible heat and latent heat distribution (Peters et al., 1998). Due to the changes of land-atmosphere heat exchange, the surface soil heat condition is bound to change. The soil temperature change is accompanied by alteration of the water condition; therefore, the changes of land surface sensible heat and latent heat values also have a certain effect on STC. Generally, the ground-air temperature difference can be used to indicate the land surface sensible heat condition. In this study we associated the changes of thermal conductivity with the ground-air temperature difference and the surface soil moisture content to analyze the effects of the changes of the latter two on the thermal conductivity. Through the selection and analysis of correlation factors, we obtained the following regression equation:

$$STC = 0.852492 - 0.0125198(T_0 - T_a) + 1.37868\theta_u \qquad [4]$$

Where T_0 is the ground surface temperature and Ta is air temperature. The explanations of other symbols are the same as described above. The multiple correlation coefficient of the equation is 0.95 and passes the 0.01 significance test. Some meteorological stations record STC and soil moisture content observations, but most meteorological stations do not record soil water measurement data. Therefore, the application of Equation [4] is limited by the availability of observation data. Our data analysis indicates that the soil moisture content in the 0–10-cm layer in the study region can be expressed by the function of water vapor pressure:

$$\theta_u = 0.1298\ln(E) + 0.0495 \qquad [5]$$

Where E is water vapor pressure (hPa).
Then Equation [4] can be rewritten as:

$$STC = 0.8525 - 0.0125(T_o - T_a) + 1.3787(0.1298\ln(E) + 0.0495) \qquad [6]$$

The performance of the above model was tested by the independent data from January 2005 to December 2013. Such result was shown in Figure 3. It can be seen from Figure 3 that the performance of the model was good. The STC observed and tested were in good agreement. Averagely, the absolute error between the STC observed and tested ranged from 0.01 to 0.200 W.m^{-1}.K^{-1}, with an average of 0.064 W.m^{-1}.K^{-1}. The relative error between them ranged from 0.1% to 18.5%, with an average of 6.4%. Our error analysis shows that Equation (6) has a small error, its data are easy to collect, and it can therefore be readily used to estimate the monthly mean STC of the active layer in the northern QTP.

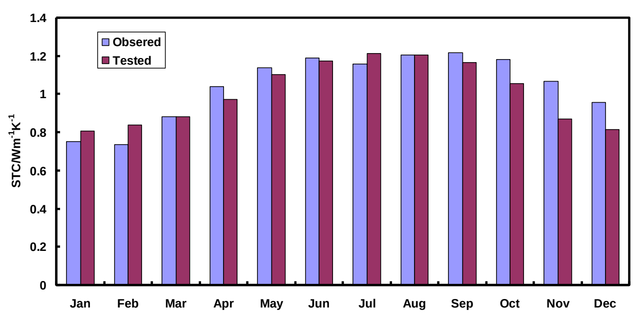

Figure 3 Comparison of observed results and tested results

4 CONCLUSIONS

From the above discussion and analysis, we have drawn the following conclusions:

1. In the study region, Ts_5 (soil temperature at the 5-cm depth) varied between 0.3 ~ –1.6 °C, Ta varied between –5.1 ~ 3.2 °C, TDDs and FDDs values increased with decreasing latitude, and at decreasing latitude the absolute value of the ratio of the two increased from 0.210 to 0.422.

2. STC in the study region was low in FS but high in UFS, exhibiting obvious seasonal variation. Thermal conductivity increased with increasing vegetation cover of the underlying surface, and the STC of alpine frost meadow soil was higher than that of alpine frost steppe soil.

3. Low STC in FS in the study area was related to the fine-grained soil structure, the small unfrozen water content, and the low saturation degree.

Soil moisture content has a significant effect on thermal conductivity. STC increases with increasing soil moisture content, and the variations of ground-air temperature difference and water vapor pressure are indicative of the monthly mean thermal conductivities of the surface layer of the active layer in the northern QTP.

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