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Synergistic effect of vegetation and air temperature changes on soil water content in alpine frost meadow soil in the permafrost region of Qinghai-Tibet

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Abstract:

Seasonal changes over 2 years (2004–2006) in soil moisture content (θ_v) of frozen alpine frost meadow soils of the Qinghai-Tibet plateau permafrost region under three different levels of vegetation cover were investigated. Vegetation cover and air temperature changes had significant effects (synergistic effect) on θ_v and its distribution in the soil profile. During periods of soil freezing or thawing, the less the vegetation cover, the quicker the temperature drop or rise of soil water, and the shorter the duration of the soil water freeze–thaw response in the active soil layer. Under 30% and 65% vegetation cover the amplitude of variation in θ_v during the freezing period was 20–26% greater than that under 93% cover, while during the thawing period, it was 1·5- to 40·5-fold greater. The freezing temperature of the surface soil layer, ${}^{f}T_s$, was 1·6 °C lower under 30% vegetation cover than under 93% vegetation cover. Changes in vegetation cover of the alpine frost meadow affected θ_v and its distribution, as well as the relationship between θ_v and soil temperature (T_s). As vegetation cover decreased, soil water circulation in the active layer increased, and the response to temperature of the water distribution across the soil profile was heightened. The quantity of transitional soil phase water at different depths significantly increased as vegetation cover decreased. The influence of vegetation cover and soil temperature distribution led to a relatively dry soil layer in the middle of the profile (0·70–0·80 m) under high vegetation cover. Alpine meadow θ_v and its pattern of distribution in the permafrost region were the result of the synergistic effect of air temperature and vegetation cover. Copyright © 2007 John Wiley & Sons, Ltd.

KEY WORDS permafrost; soil water content; alpine frost meadow; vegetation cover; temperature change

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INTRODUCTION

Global climate changes have altered natural ecosystems significantly in many regions of the world. Such alterations have included shifts in plant community structure and composition, biological productivity and biodiversity (Walker and weller, 1991, Bubier *et al*., 1999; McGuire, 2002). As the global climate warms, glaciers and frozen soils in some sensitive regions are altered significantly, thereby accelerating the degradation of alpine ecosystems (Jorgenson *et al*., 2001; Walker *et al*., 2003). Transects within the cryosphere of arctic regions have shown that alpine ecosystems are quite sensitive to global climate changes. Alterations to such ecosystems can lead to dramatic changes in soil physical properties, soil- and surface-water dynamics and in the soil carbon cycle. This, in turn, exerts a profound influence over the entire biosphere (Weller *et al*., 1995; Jorgenson *et al*., 2001; Christensen *et al*., 2004). Given the magnitude of a frost region's heat sink, such a region's terrestrial ecosystems strongly influence the water cycle and heat balances of the regional land–atmosphere systems (Carey and Woo, 1999; Rouse, 2000). Consequently, the energy and water balance of the Qinghai-Tibet plateau has an important influence on the Asian monsoon system, and represents an important component of the global climate's energy and water cycles (Zhang *et al*., 2003).

The permafrost region, located in the interior of the Qinghai-Tibet Plateau, represents a distinct cryospheric environment, housing a number of typical alpine landscapes, including alpine meadows and alpine steppes, whose ecosystems and water cycles have clearly been adversely affected by global climate changes (Li *et al*., 1998; Wang *et al*., 2001). Therefore, an understanding of the laws governing variations in θ_v and associated heat of alpine frost ecosystems is important in understanding regional variations in water cycling arising from global climate changes (Rouse, 2000; Zhang *et al*., 2003). The implementation of the GAME/Tibet study in the Qinghai-Tibet plateau seeks to deepen our understanding of issues such as water and heat fluxes, regional water and heat balances and changes in climatic conditions in the region (Liu *et al*., 2003; Zhang *et al*., 2003). However, given the Qinghai-Tibet Plateau permafrost region's high sensitivity to climatic

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changes, the vegetation–soil–atmosphere water cycle and heating–cooling processes coupled thereto are quite complex. Some early studies in this region have shown that there are close ties between permafrost and the extent of vegetation; indeed, permafrost degradation was found to significantly alter the vegetation (Zhao *et al*., 2000; Wang *et al*., 2006). Consequently, one must address the important question of what changes in surface soil water processes will occur under the combined effects of the factors including climate, permafrost and vegetation. The main purpose of this study was to analyse changes in soil water content and profile distribution that occurred as the result of the degradation in vegetation typical of the permafrost region.

STUDY AREA AND METHOD

Study area

Located in the upper watershed of the Zuomao Kong River, a tributary of the Yangtze River, the Fenghuoshan permafrost region of the Qinghai-Tibet plateau (92°50′–93°3′E and 34°40′–34°48′N) served as the study area. It embraces a total area of 127·63 km^2 at elevations ranging from 4510 m to 4723 m (Figure 1). The annual mean (1973–2005) air temperature (T_a), relative humidity and precipitation were −5·2 °C, 57%, and 310·7 mm, respectively. The vegetation is dominated by *Kobresia pygmaea* C.B. Clarke and *Kobresia humilis* Serg. According to the degree of degradation, vegetation cover of the region's grasslands was divided into three categories: non-degraded, moderately degraded and severely degraded, corresponding to 93%, 65% and 30% coverage. In severely degraded grassland *Kobresia sp.* was replaced by *Festuca sp.* and *Poa spp.* (Wang *et al*., 2001; Zhou, 2001).

Soil in the region is dominated by mattic gelic cambisols. The main physical properties and nutrient contents of the region's alpine frost meadow soils under different levels of cover are presented in Table I. In severely degraded alpine frost meadow, the coarse sand and gravel contents of the topsoil were significantly increased. The greater the vegetation cover, the larger the soil organic matter content and the lesser the soil bulk density (Table I). In the study area the permafrost was well developed, averaging between 50 and 120 m in depth, with an active layer of 0·8–1·5 m (Zhou *et al*., 2000). There are two meteorological stations, and the meteorological indexes, such as air temperature, land surface temperature, precipitation, wind, humidity, total radiation, net radiation were observed.

Periods of soil water changes in alpine frost meadows

As shown in Figure 2, the coupled changes in soil moisture content (θ_v) and soil temperature (T_s) occurring over the year in the active layer of permafrost soils can be clearly divided into four periods:

- Freeze initiation period (FIP, the period A): From 15–20 October, when the active layer begins freezing, to 10–15 November when it is entirely frozen. This entails a decrease in θ_v from 25–35% to 1·4–13·5%.

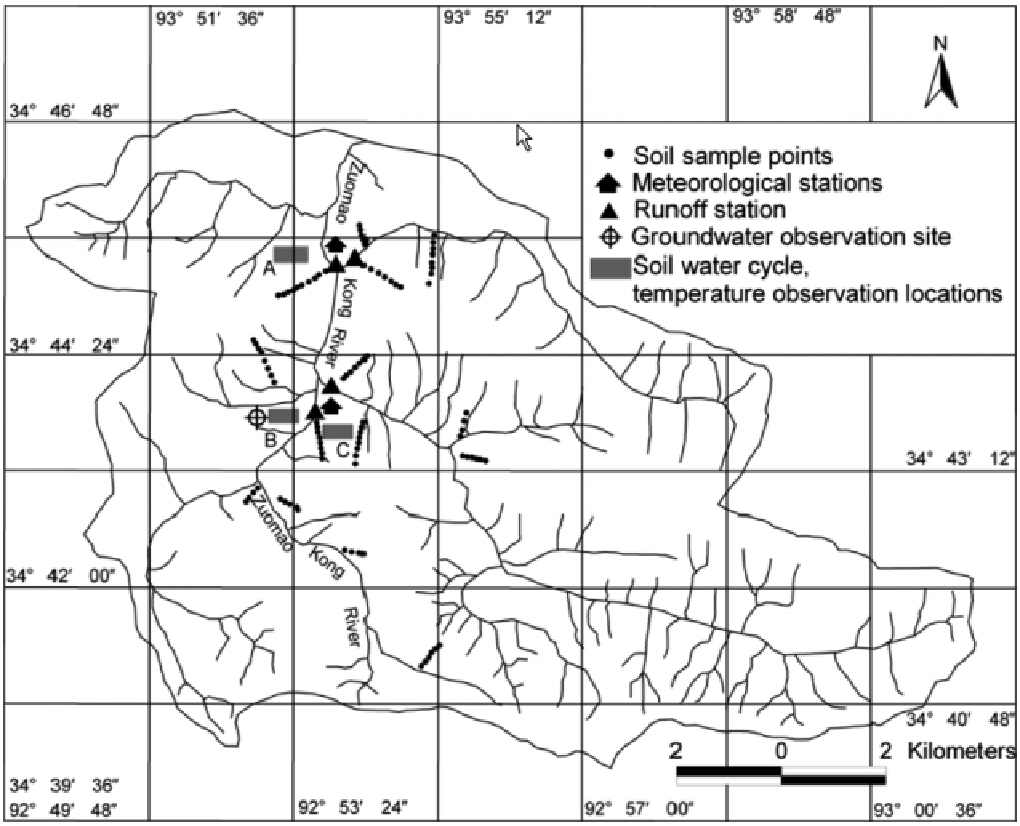

Figure 1. Location of the study area

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Table I. Physical and chemical characteristic of selected alpine frost meadow soils of the Qinghai-Tibet Plateau

Vegetation type		Permafrost				Granularity			
Type	Cover (%)	Depth (m)	Active layer (m)	Soil Profile (m)	Bulk density (Mg m^{-3})	>0·5 mm (%)	<0·1 mm (%)	Organic matter (%)	Salt content (%)
Festuca sp., Poa spp.	30	50–120	1·0–1·5	0–0·10	1·26	0·11	93·1	0·66	0·27
				0·10–0·20	1·34	0·09	85·4	0·25	0·16
				0·20–0·40	1·51	0·07	61·2	1·13	0·04
Kobresia humilis	65	50–120	0·8–1·5	0–0·10	1·10	0·10	93·2	1·30	0·32
				0·10–0·20	1·21	0·09	92·9	0·80	0·19
				0·20–0·40	1·35	0·09	72·4	0·77	0·13
Kobresia humilis	93	50–120	0·8–1·5	0–0·10	0·95	0·09	93·3	1·10	0·26
				0·10–0·20	1·09	0·08	93·6	1·63	0·25
				0·20–0·40	1·29	0·10	61·8	1·38	0·17

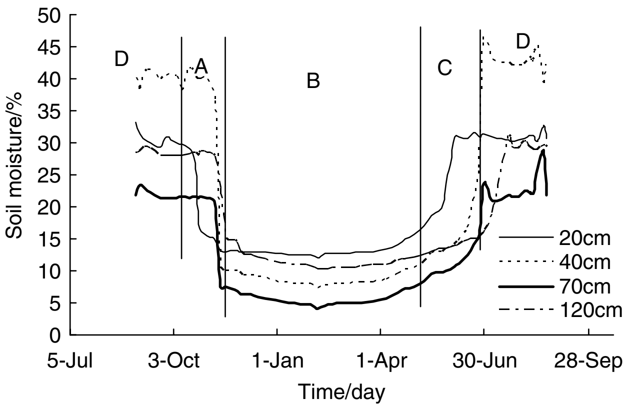

Figure 2. Changes in soil water content, and its separation into fully frozen, thawing, fully thawed, and freezing periods

- Entirely frozen period (EFP, the period B): From 15–20 November to the last 10 days of April, the active layer of the soil is in a fully frozen state and θ_v remains at a constant level between 0·3 and 11·0%.
- Thaw initiation period (TIP, the period C): From the first 10 days of May to the last 10 days of June, the active layer of the soil (0–1·5 m) is thawing, and θ_v gradually increases from 0·3–11·0% to the peak value of the early summer (26–43% in the upper 0·40 m of the soil).
- Entirely thawed period (ETP, the period D): From early July to the first 10 days of October, the entire active layer (0–1·50 m) is in a thawed state.

The layout of soil water observational experiments

Three soil water–temperature coupling observation locations A, B and C were selected within the Zuomao Kong watershed, with locations A and C in the alpine frost meadow grassland and location B in the alpine frost swamp grassland (Figure 1). The paper is based entirely on data from site A. Three 20 m × 5 m (slope length × width) sampling plots, with slopes of 18–21° and vegetation covers of 30%, 65% and 93%, were selected within location A. Two soil moisture observation points were positioned one-third and two-thirds of the way down the slope of each plot, each housing a pair of 1·50 m deep soil water and temperature observation wells. In each well, five soil moisture sensors and soil temperature (T_s) sensors were located at depths 0·20, 0·40, 0·70, 1·20 and 1·50 m. Before burying the water probes and temperature sensors, the soil bulk density at each point was determined by the cutting ring method and the soil particle-size composition was also determined.

θ_v was determined by frequency domain reflectometry (FDR), using a calibrated soil moisture sensor equipped with a Theta-probe (Holland Eijkelamp Co. Giesbeek, Nerthlands). θ_v was derived from changes in the soil's dielectric constant, converted to a millivolt signal, the accuracy being ±2%.

T_s was monitored using a thermal resistance sensor sensitive to temperature changes in the range of −40 °C to 50 °C, with an overall system precision of ±0·02 °C. The thermal resistance sensors were developed by the State Key Laboratory of Frozen Soil Engineering using digital multimeters (Fluke 180 series, Fluke Co. USA), and had been successfully used in the Qinghai-Tibet plateau over the previous 20 years (Wu *et al.* 2002, 2004).

Both θ_v and T_s were monitored simultaneously at 2 h intervals from April to November and at 6 h intervals from December to March. Several portable micrometeorological stations were set up in the experiment to measure the climatic factors T_a (at 1·2 m height), precipitation, wind velocity and direction, and radiation.

Analysis of θ_v and profile distribution changes

To investigate the quantitative and spatial response of θ_v to air temperature and changes in vegetation cover objectively during the permafrost region's freeze–thaw cycles, a number of indexes were selected:

- The thaw-rise time t_s represents the time at the inflexion point in the plot of θ_v versus time, i.e. where θ_v begins to increase following the initiation of thawing (Figure 2, Section C).
- The thaw-rise duration t_r represents the time between soil water beginning to thaw and reaching the peak early summer values (Figure 2, Section C).
- The soil water thaw-rise amplitude ΔW_s and the soil water thaw-rise rate V_s:

$$\Delta W_s = (\theta_m - \theta_0)/\theta_0 \quad (1)$$

$$V_s = (\theta_m - \theta_0)/t_r \tag{2}$$

where, θ_m and θ_0 are the soil moisture content at the thaw-rise peak value and at its initial value, respectively.

For the freezing period (Figure 2, Section A) similar indexes describing the dynamics of soil water were developed: the freeze-fall time of soil water t_d, the freeze-fall duration t_z, the soil water freeze-fall amplitude ΔW_d and the freeze-fall rate V_d. These were used to describe the water profile distribution changes during freezing periods.

Analysis of changes in $\theta_v - T_s$ coupling and amount of phase transitional water

Soil temperature parameters including soil freezing temperature (${}^{f}T_s$), soil thawing temperature (${}^{t}T_s$), soil temperature above freezing ($\Delta T_{>f} = T_s - {}^{f}T_s$), and, Q, the quantity (kg) of phase transitional water per unit area of a soil column of depth h, were selected to analyse soil water dynamics in permafrost soils under different levels of vegetation cover. The parameters ${}^{t}T_s$ and ${}^{f}T_s$ were determined by the inflexion point method, based on plots of θ_v versus T_s. As T_s decreased, the point where θ_v dropped sharply then remain relatively stable, was taken as ${}^{f}T_s$. At this point the simplified assumption was made that all water was frozen except a small residual quantity of unfrozen water. Conversely, as the T_s of a frozen soil rose, the point at which water content shot up sharply, to then remain relatively stable, was taken as ${}^{t}T_s$.

When $T_s > 0\,^{\circ}\mathrm{C}$, and when $T_s < 0^{\circ}$ but soil water was not yet frozen, θ_v calculated from FDR data represented the volumetric soil liquid-water content. Barring significant shifts in FDR data, θ_v immediately before the soil froze was taken as the volumetric soil solid-water content (Zhou *et al.*, 2000; Zhang *et al.*, 2003). The same method was applied in the permafrost region of eastern Siberia (Sugimoto *et al.*, 2001).

As shown in Figure 3, the precipitation in the study region falls mainly in the three months from July to September, accounting for 83% of the annual total precipitation. During the freezing season from November to April, precipitation was less than 5 mm. The distribution of yearly air temperature has an obvious consistency with the precipitation processes, while the peak temperature occurs from July to August and the monthly average air temperature are all lower than 0 °C. During the FIP and TIP periods, the snowmelt (discussed again later) and rain infiltration was negligible. The value of Q in freezing and thawing soil was determined by assuming that θ_v determined by FDR during the entirely frozen period represented the residual amount of water. During the soil freezing period the extent of soil water phase transition (liquid → solid) can be expressed as (Zhou *et al.*, 2000; Jansson and Karlberg, 2001) and represents the possible amount of liquid water present:

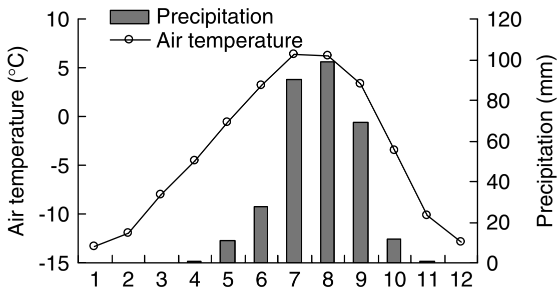

Figure 3. Distribution pattern of monthly precipitation and air temperature

$$Q = (\overline{\theta_m} - \overline{\theta_0})\rho h \tag{3}$$

where $\overline{\theta_m}$, $\overline{\theta_0}$ represent the mean initial and final remaining freeze-fall θ_v to a depth h (m%), and ρ is soil bulk density (Mg m^{-3}).

Similarly, if $\overline{\theta_m}$ represents the mean initial soil water thaw-rise value, Equation (3) can be used to calculate the quantity of phase transitional water Q_{ir} (solid → liquid) during the soil thawing period.

Regression models (Table II) were used to analyse $\theta_v - T_s$ relationships for both the freezing and thawing periods, as occurred under different levels of vegetation cover.

RESULTS AND DISCUSSION

Effect of vegetation cover changes on soil water during the thawing period

From the last 10 days of April when mean daytime T_a in the study area began to rise above 0 °C, to early July, when it rose above 5 °C and the frozen soils were thawed to a depth of 1·5 m, soil water experienced a redistribution in the profile with changes in freeze-thaw depth. As shown in Table II, the net radiation and surface heat flux entering the soil were different under different levels of vegetation cover. The greater the vegetation cover, the larger the net radiation and lower the heat flux into the soil. In company with the frost meadow degradation, the soil structure

Table II. Average radiation and surface soil heat flux in different time under different vegetation cover (MJ m^{-2} day^{-1})

Vegetation cover (%)	30			65			93		
Periods(month)	5–6	7–8	9–10	5–6	7–8	9–10	5–6	7–8	9–10
Net radiation	7·4	9·9	6·2	8·2	11	6·7	8·8	12·2	7·2
Sensible heat	5	4·4	5·2	4·8	4·2	4·9	4·3	3·9	4·5
Latent heat	1·4	3·6	1·4	2·7	5·3	2·1	4	7·1	2·9
Surface heat flux	1	1·9	−0·4	0·7	1·5	−0·3	0·5	1·2	−0·2

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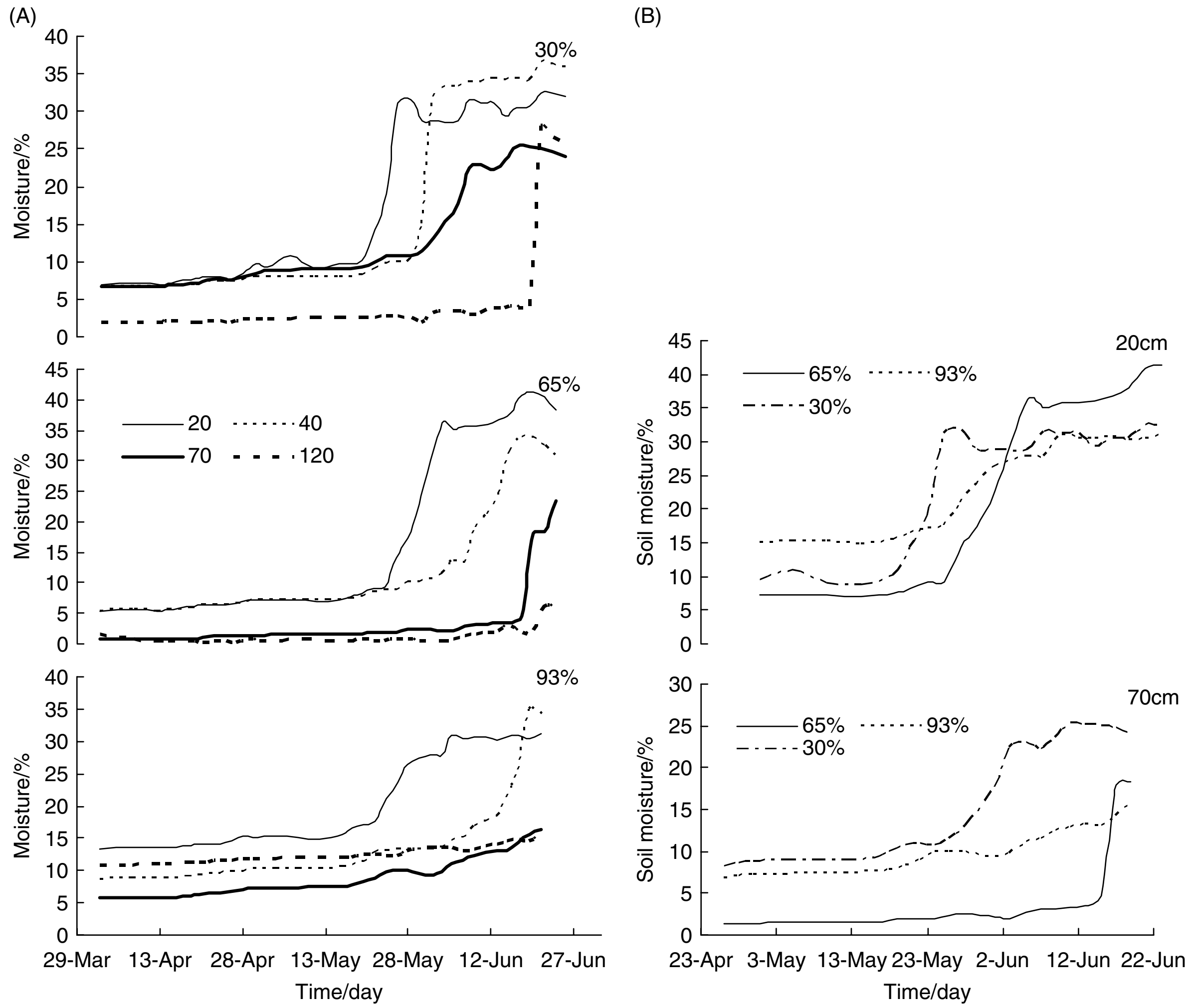

Figure 4. Soil water content and dynamics of an alpine frost meadow grassland soil profile in the permafrost region of the Qinghai-Tibet plateau during the thawing period. (A) for a specific soil layer (B) under different levels of vegetation cover

of the top layer changed as shown in Table I. As a result, θ_v and its distribution through the profile changed significantly under different levels of vegetation cover (Figure 4).

Prior to complete thawing (early April to mid-May), when θ_v and its distribution in the profile was locked to the freezing period's pattern, small differences existed between the different levels of vegetation cover (Figure 4A). This was mainly manifested under 30% cover, where θ_v in the 0–0·70 m soil layer was initially greater than in deeper layers, although this difference gradually disappeared by the end of April, at which point changes of θ_v in the deep layers became less evident. Under 65% cover, θ_v in the 0–0·40 m soil layer was greater than in deeper layers, where it gradually rose with time, until early May when the change in θ_v below 0·70 m became less evident. Under 93% cover, θ_v distribution in the profile differed greatly from that of soils with less cover: θ_v was greatest in the 0–0·20 m soil layer and in soil layers $\approx$1·20 m in depth, whereas θ_v at a depth of 0·70 m was lower, only catching up to topsoil levels in early May.

From 20 May to 20 June, when the active layer (0–1·20 m) was entirely thawed, θ_v varied dramatically, rising to peak values in the early summer (Figure 4A, Table III). The response time of θ_v to increased T_a

Table III. Response of thaw-rise time (t_s) and duration (t_r), and soil moisture thaw-rise amplitude (ΔW_s) for different soil depths and levels of vegetation cover of alpine frost meadow soils of the Qinghai-Tibet Plateau

Depth	0·20 m			0·40 m			0·70 m			1·20 m		
Soil cover	t_s d/m	t_r days	ΔW_s	t_s d/m	t_r days	ΔW_s	t_s d/m	t_r days	ΔW_s	t_s d/m	t_r days	ΔW_s
65%	22/5	8	2·89	28/5	17	2·39	15/6	9	2·98	21/6	3	2·10
93%	21/5	14	0·81	3/6	16	1·55	18/6	16	0·36	3/7	10	0·14
30%	16/5	5	2·51	24/5	7	2·34	28/5	12	1·21	14/6	2	5·67

showed a significant difference with level of vegetation cover.

Under 30% cover, θ_v in the 0–0·20 m soil layer began its thaw-rise in mid-May (t_s), and between this date and the end of the month θ_v began its thaw-rise at a depth of 0·70 m. Initiation of the thaw-rise in θ_v at a depth of 1·20 m lagged about 1 month behind that of the surface soil layer. Similar trends in θ_v were apparent for the 93% and 65% cover soils.

At a given soil depth t_s varied significantly with the level of cover (Figure 4B and Table III). The t_s of the 0–0·20 m soil layer occurred 6–7 days earlier under 30% cover than under 65% cover, and 5–6 days earlier than under 93% cover. At depths of 0·40 and 0·70 m, respectively, the t_s of permafrost soil under 30% cover was 4–5 days and 15–18 days earlier than under 65% cover, and 9–10 days and 19–21 days earlier than under 93% cover. At a depth of 1·20 m, t_s under 93% vegetation cover lagged 17–19 days and 11–14 days, respectively, behind that of soils under 30% or 65% cover. Clearly, the greater the vegetation cover, the later the soil water thaw-rise started.

At given depths (0·20 m, 0·40 m, 1·20 m), the duration of the alpine frost meadow soil's water thaw-rise response (t_r) was 8–9 days longer under a vegetation cover of 93% than under a cover of 30% (Table III). The relative amplitude of thaw-rise in θ_v (ΔW_s) reflects the degree of variation of θ_v. Under 93% cover ΔW_s was 1·5- to 40·5-fold lower than under 30% cover. At a depth of 1·20 m, the ΔW_s under 30% cover was 2·7- and 40·5-fold greater than under covers of 65% or 93%, respectively. These differences in t_r and ΔW_s led to significant differences in the distribution of θ_v through the soil profile under different levels of vegetation cover, during the thawing period. As the alpine frost meadow became degraded and vegetation cover declined, the response of θ_v to T_a became more pronounced, t_s was earlier and ΔW_s increased. Thus, the greater the vegetation cover, the longer t_r and the lower ΔW_s (Figure 4B).

Under 30%, 65% and 93% vegetation cover, respectively, the water thaw-rise rate of alpine frost meadow soils (V_s) varied between 1·15–11·9% d^{-1}, 1·4–3·36% d^{-1}, and 0·2–1·29% d^{-1} across the 0–1·2 m soil profile, being greatest at depths of 1·20 m, 0·2 m, and 0·4 m (Table IV). Thus, the lower the vegetation cover of the alpine frost meadow soil, the greater V_s in the active layer, and the greater its variation between different depths. During the thawing periods (TIP, period C in Figure 2), precipitation was very little and the effects of rain infiltration on the water distribution in the soil profile was negligible. The differences in heat flux, heat consumption level and ice content in different soil layers were the main drivers of differences in water phase change and water distribution in the soil profiles under the different levels of vegetation cover. Generally, the greater the vegetation cover, the lower the heat flux entering the subsurface soil (Table II), and the lower the thawing rate.

Distribution of soil water under different vegetation covers during the freezing period

During the freezing period the response of θ_v to T_s differed according to the level of vegetation cover (Figure 5). During the freezing process, the 0–0·20 m soil layer under 30% cover began its freeze-drop (t_d) in early October, 9 and 14 days earlier than under 65% and 93% vegetation, respectively. The greater the cover, the later the t_d of the active layer of the soil profile (Table V). As with t_s for the thawing process, t_d was delayed by an increase in soil depth. At a depth range of 0·70–1·20 m, the difference in t_d between vegetation covers of 30% and 65% was 7–11 days, whereas between 30% and 93% cover it was 12–23 days. These phenomena are consistent with the soil heat conductivity and temperature conductivity under different levels of vegetation cover. The greater the cover, the larger the soil organic matter and lower the soil bulk density, and the lower the heat and temperature conductivity. As a result, soil freezing rate was greater under low vegetation cover than that under high vegetation cover (Zhou *et al.*, 2000).

The freezing duration (t_z) of the 0–0·20 m soil layer under different vegetation covers was roughly the same (Table IV, Figure 5), but 1·8- to 5-fold longer than the thawing duration (t_r). Below 0·40 m in depth, t_z under 93% cover was 4–12 days and 7–13 days shorter than under 65% or 30% cover, respectively. Beginning with

Table IV. Rate of soil thaw-rise and freeze-drop (% day^{-1}) in soil moisture content (V_s and V_d, respectively) at difference depths and under different vegetation covers, for alpine frost meadow soils of the Qinghai-Tibet Plateau

Vegetation cover	0·20 m		0·40 m		0·70 m		1·20 m	
	V_s	V_d	V_s	V_d	V_s	V_d	V_s	V_d
65%	3·36	1·01	1·41	0·93	1·52	1·19	1·40	0·63
93%	0·96	0·59	1·29	2·40	0·37	1·74	0·20	1·15
30%	4·48	1·07	3·44	1·41	1·15	0·74	7·90	1·48

Table V. Response of soil water freeze-rise characteristic parameters for different soil depths and levels of vegetation cover of alpine frost meadow soils of the Qinghai-Tibet Plateau

Depth	20 cm			40 cm			70 cm			120 cm		
Soil cover	t_d (d/m)	t_z days	ΔW_d	t_d (d/m)	t_z days	ΔW_d	t_d (d/m)	t_z days	ΔW_d	t_d (d/m)	t_z days	ΔW_d
65%	15/10	26	0·83	29/10	16	0·71	1/11	11	0·85	3/11	26	0·85
93%	20/10	25	0·52	5/11	12	0·75	7/11	8	0·67	15/11	14	0·59
30%	6/10	23	0·77	14/10	19	0·79	24/10	21	0·68	22/10	26	0·89

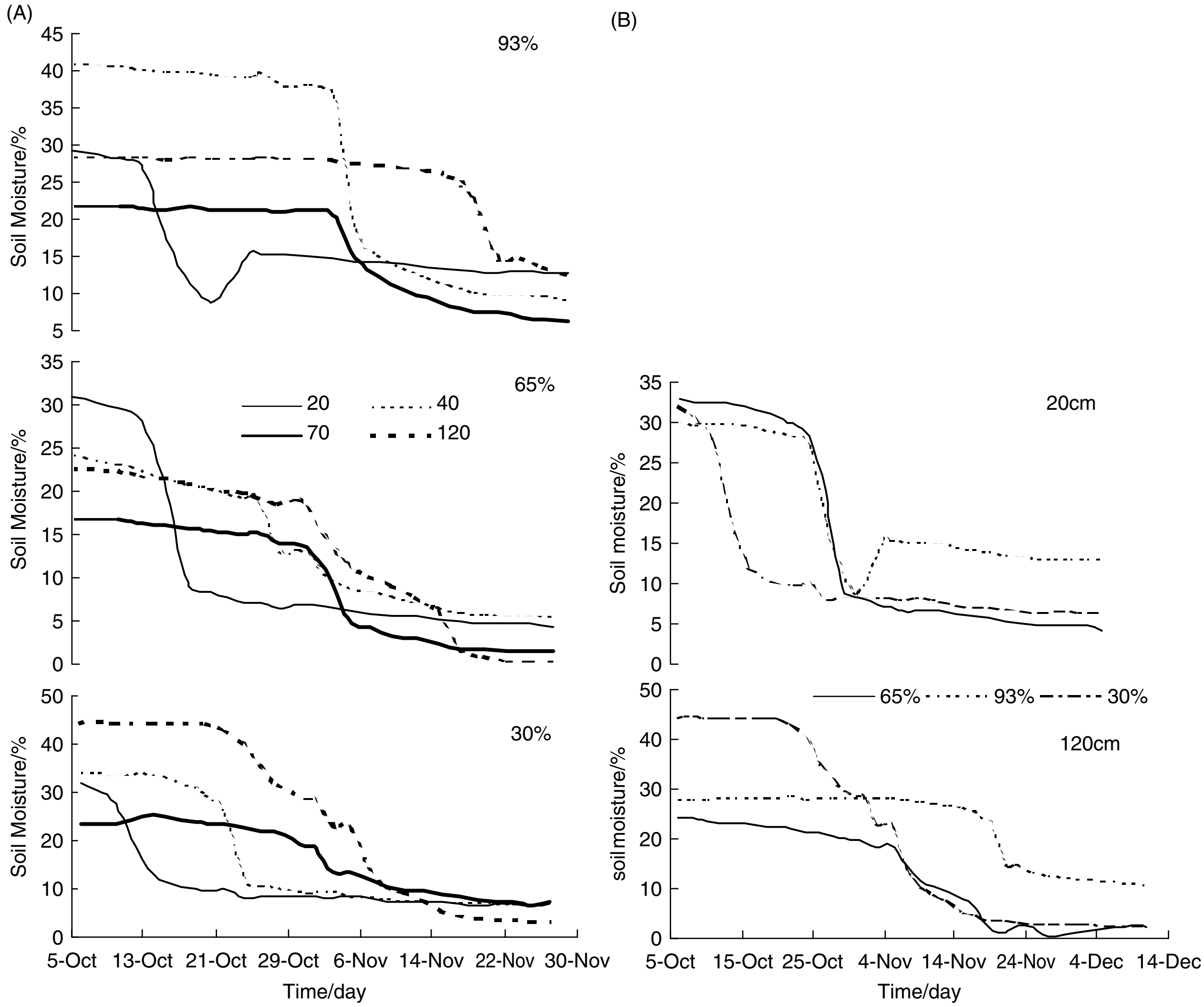

Figure 5. Soil water content and distribution within the active soil profile of an alpine frost meadow grassland soil of the Qinghai-Tibet plateau under different levels of vegetation cover during the (A) thawing period and (B) freezing period

water freezing in the surface soil layer, to it freezing at a depth of 1·20 m, t_z values for the entire profile (i.e. from when the surface soil layer froze to when the soil at the 1·20 m depth froze) were 15, 19, and 25 days under vegetation covers of 30%, 65%, and 93%, respectively. Thus, the lower the vegetation cover, the longer the t_z at a given depth and the shorter the t_z over the whole active layer, which is the opposite of the thawing process. The freeze-fall amplitude of soil water (ΔW_d) was similar to that for the thawing process. For a given soil depth, the greater the vegetation cover, the lower the ΔW_d.

During the freezing period the response of the alpine frost meadow soil water freeze-fall rate (V_d) to the level of vegetation cover was complex. Under 30% vegetation cover, V_d for the 0–0·20 m soil layer was 30% greater than under 65% or 93% cover (Table IV). This was similar to changes in the thaw-rise rate (V_s). However, in the 0·40–0·70 m depth range V_d was greater under 93% cover than under 30% cover, the opposite of what was found for V_s.

Soil water content and profile distribution in the fully thawed and frozen periods

During August and September, when the study area's 1·20 m deep active layer was entirely in a thawed state, θ_v of the 0·20–0·40 m soil layer under a vegetation cover of 93%, significantly exceeded that of other soil layers under the same cover (Figure 6A). Under 65% cover θ_v differed significantly across the active layer profile, with slightly greater values in the surface soil layer, and low values in the 0·40–0·70 m layer. Under 30% cover, θ_v at 1·20 m depth was significantly greater than that in other layers, while θ_v was lowest at 0·70 m depth. In the study region, more than 83% of annual total precipitation falls during the fully thawed periods, and the infiltration water has an obvious effect on the soil water content during this period. The soil saturated hydraulic conductivity of the 0·10–0·30 m soil layer was significantly greater than that in other layers (especially under 93% vegetation cover) (Figure 7), and the coagulative water caused by the large temperature difference between day and night during thawing season

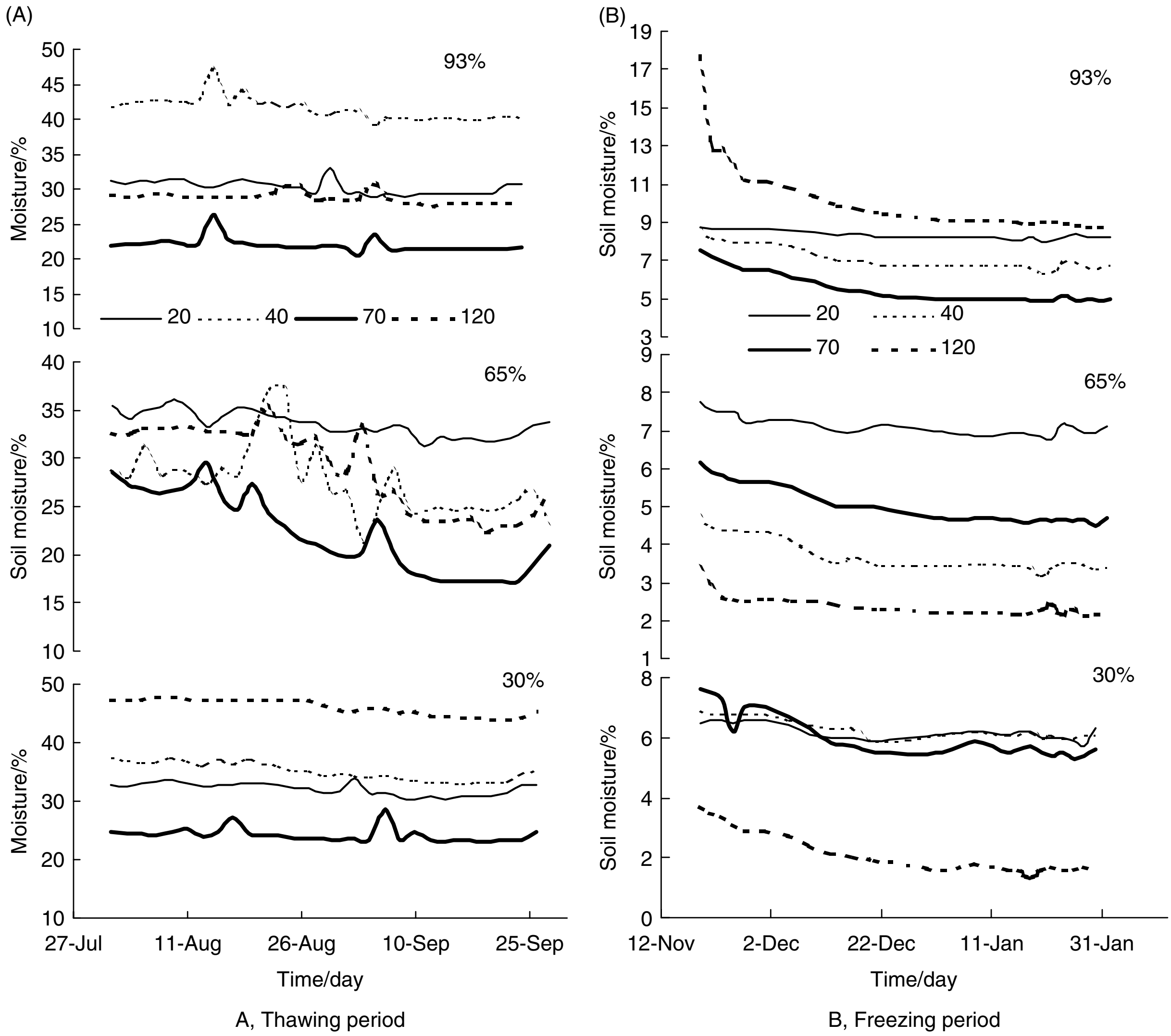

Figure 6. Soil water content and dynamics of an alpine frost meadow grassland soil profile in the permafrost region of the Qinghai-Tibet plateau during the freezing period. (A) for a specific soil layer (B) under different levels of vegetation cover

was concentrated in the top soil layer (Zhou *et al*., 2000; Zhao *et al*., 2000), which resulted in high water content in the upper profile during thawing periods. For the thawing depth achieved to 1·2–1·5m during August and September, the migration of soil water to the freezing fronts caused the water content at 1·2 m soil depth to be relatively higher than that at mid-profile. θ_v showed an upward build-up and downward movement in the profile, clearly leaving a drier mid-profile (0·70–0·80 m), where $\theta_v < 25\%$ (Figure 6A).

From late November to the subsequent March, when the study area's 1·20 m deep active layer was entirely in a frozen state, the θ_v determined by FDR was low (1·3–10%), the greater portion of soil water existing in the form of solid ice. The soil water profile distribution in the frozen period was quite different from that in the thawed period (Figure 6B). Regardless of the level of vegetation cover, the 0–0·20 m surface soil layer generally bore a greater quantity of frozen water. Under 93% cover, θ_v was at its greatest in the deeper layer (1·20 m depth), but low in the mid-depth range 0·40–0·70 m, indicating that soil water was clearly concentrated in the upper and lower layers of the soil profile over the entire frozen period. However, under less than 65% cover, θ_v was lowest at a depth of 1·20 m (Figure 6B). With a decrease in vegetation cover, θ_v of the middle layers gradually increased, while θ_v in the lower layers decreased.

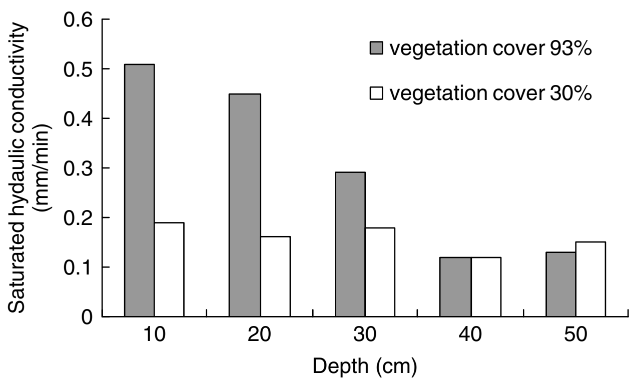

Figure 7. Soil saturated hydraulic conductivity at difference depth layer under vegetation level of 93% and 30%

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Soil water–heat coupling changes under different vegetation covers

In the study region, because the soil salt content was relative low under different vegetation cover and differed only slightly (Table I), the effect of soil salt content on the soil water–heat regime is not relevant in this study. The soil water content and soil texture were tied up to the level of vegetation cover, as detailed above (Table I and Figures 4 and 5), and soil texture and vegetation cover strongly modify the soil heat flux, latent heat flux and sensible heat flux profile (Table II). Therefore, soil water–heat coupling changes were interrelated tightly with the vegetation cover.

(1) *Impacts of vegetation cover changes on soil water freezing temperature*. Generally, a value of $T_s = 0\,°C$ is used as the freezing indicator for soils in the permafrost region. However, a number of studies have shown that the soil freezing temperature (${}^{f}T_s$) is related to a number of factors including θ_v, salt content and soil texture, such that different ${}^{f}T_s$ values may exist in different regions (Zhou *et al*., 2000). In fact, an important indicator of soil freezing is that the gravity water is frozen (Zhou *et al*., 2000; Zhang *et al*., 2003), so that if soil water in a given layer is in the B section in Figure 2, the soil in that layer is considered to be frozen.

Inflexion points in soil water–heat coupling plots (i.e. θ_v versus T_s; Figure 8) for different depths and levels of cover permitted the determination of ${}^{f}T_s$ under different conditions (Table VI). At a given depth in the soil profile, the lower the vegetation cover, the lower the ${}^{f}T_s$. In the 0–0·20 m soil layer, ${}^{f}T_s$ under 65% and 30% cover was 0·68 °C and 1·67 °C lower, respectively, than under 93% cover. At 0·70 m depth, ${}^{f}T_s$ under 65% and 30% cover was 0·1 °C and 1·23 °C lower, respectively, than under 93% cover. As the depth further increased to 1·20 m depth and beyond, ${}^{f}T_s$ gradually leveled out, and the influence of vegetation cover on ${}^{f}T_s$ gradually declined. Under a given level of vegetation cover ${}^{f}T_s$ increased with depth. The bore observation data of freezing soil temperature in the study region shown that the average freezing temperature of active soil falls in the range 0·0 °C to −3·0 °C (Zhou *et al*., 2000; Wu and Liu, 2004), which indicates that the ${}^{f}T_s$ obtained from this study were reliable.

Table VI. Soil freezing temperature (${}^{f}T_s$) and quantity of phase transitional water per unit thickness (Q) at different soil depths and under different levels of cover on alpine frost meadow soils of the Qinghai-Tibet Plateau

Soil cover	93%		65%		30%	
Soil depth (m)	${}^{f}T_s$ (°C)	Q (kg)	${}^{f}T_s$ (°C)	Q (kg)	${}^{f}T_s$ (°C)	Q (kg)
0.20	−0·10	1·40	−0·78	2·78	−1·77	3·10
0.40	−1·55	2·37	−1·81	2·46	−1·89	3·40
0.70	−0·65	2·71	−0·75	1·80	−1·88	3·19
1.20	−0·03	1·97	−0·04	1·99	−0·13	3·75

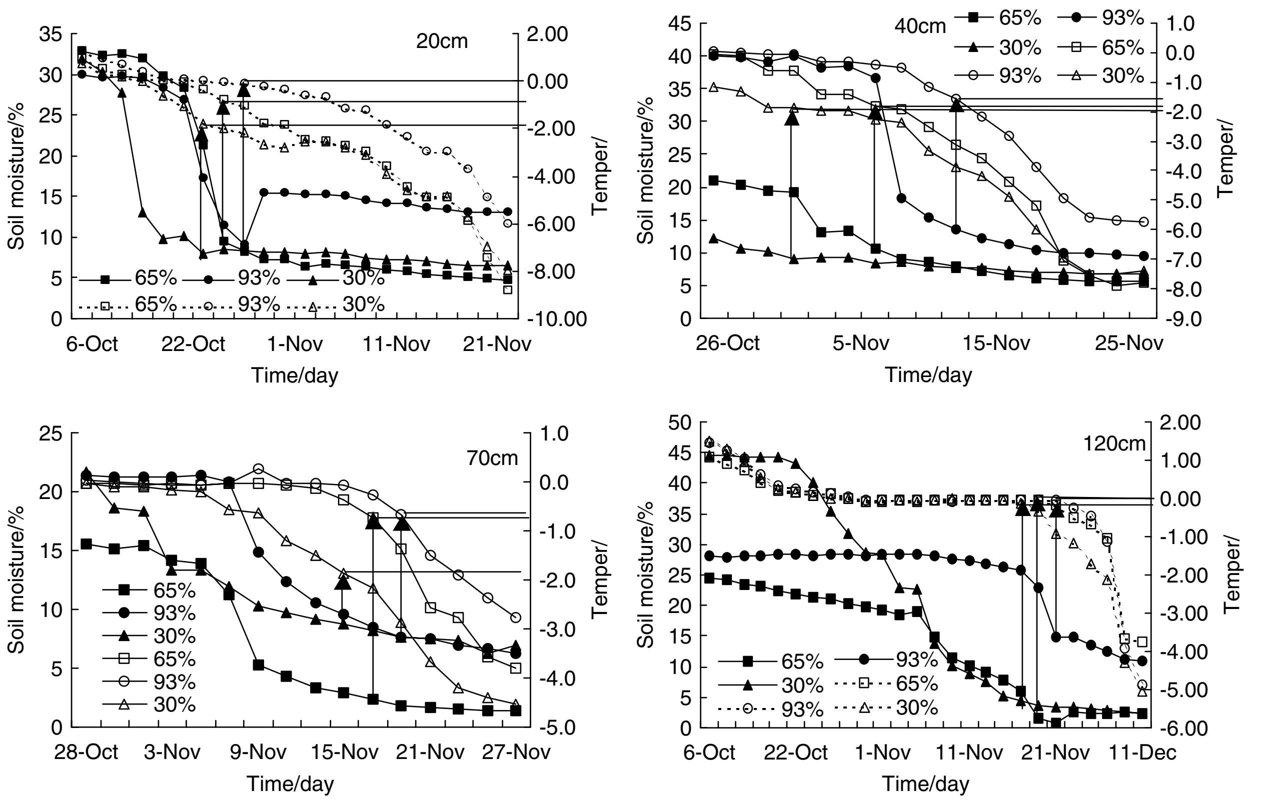

Figure 8. Soil water inflexion point and soil freezing temperature under different levels vegetation covers of an alpine frost meadow grassland soil of the Qinghai-Tibet plateau

Table VII. Soil thawing temperature (${}^{t}T_s$) and quantity of phase transitional water per unit thickness (Q) at different soil depths and under different levels of cover on alpine frost meadow soils of the Qinghai-Tibet Plateau

Soil cover	93%		65%		30%	
Soil depth (m)	${}^{t}T_s$ (°C)	Q (kg)	${}^{t}T_s$ (°C)	Q (kg)	${}^{t}T_s$ (°C)	Q (kg)
0.20	0·04	1·28	0·3	2·85	1·74	2·83
0.40	0·37	1·85	0·62	3·03	1·61	3·08
0.70	0·65	1·68	0·94	2·42	1·10	2·86
1.20	0·31	0·51	0·76	1·21	0·82	2·62

(2) *Impacts of vegetation cover changes on soil water thawing temperature*. There was no effective rain or snowmelt water infiltration to influence the soil water regime during the thawing periods in the study region. Employing the same inflexion point method as for ${}^{f}T_s$ determination, the soil water thawing temperatures (${}^{t}T_s$) under different levels of cover and at different soil depths were determined (Table VII). For a given soil profile depth, the lower the vegetation cover, the greater ${}^{t}T_s$, the converse of what occurred with ${}^{f}T_s$. Under 30% vegetation cover, ${}^{t}T_s$ of the 0–0·20 m soil layer was 1·44 °C and 1·7 °C greater than under 65% and 90% cover, respectively. At a depth of 1·20 m, ${}^{t}T_s$ under 30% cover was 0·06 °C and 0·51 °C greater than under vegetation covers of 65% or 93%, respectively. Contrary to the soil freezing process, ${}^{t}T_s$ increased with depth under a given vegetation cover.

(3) *Effect of vegetation cover changes on the soil water-heat coupling relation and the quantity of soil phase transitional water*. In the permafrost region, T_s is one of the main factors affecting soil water distribution, and θ_v and its associated heat are more closely linked than in other regions (Zhao *et al*., 2000; Jansson and Karlberg, 2001). However, how the level of vegetation cover affects the $\theta_v - T_s$ coupling relationship is not fully understood. Plots are shown of the $\theta_v - T_s$ relationships during soil freezing (Figure 9A) and thawing processes (Figure 9B) in the 0–0·20 m soil layer under different levels of vegetation cover. When the soil temperature difference ($\Delta T_{>f} = T_s - {}^{f}T_s$) was less than 5 °C, a significant ($R^2 > 0·9$) cubic relationship existed between θ_v and T_s regardless of the level of vegetation cover and under both freezing and thawing processes.

The effect of vegetation cover on the $\theta_v - T_s$ relationship was mainly manifested in three aspects. (i) During the freezing period, under the same soil $\Delta T_{>f}$, V_d decreased with increasing vegetation cover, while at a given T_s, θ_v increased with increasing cover. When $\Delta T_{>f} < -4·5$ °C, the soil was frozen and V_d remained roughly the same as T_s decreased. (ii) During the thawing period the soil $\Delta T_{>f}$ increased with decreasing vegetation cover at a given θ_v. Under vegetation cover of 93%, θ_v reached an equilibrium point (i.e. θ_v remained roughly the same as T_s increased) when $\Delta T_{>f} > 1·5$ °C, while for 65% and 30% cover, this point was reached when $\Delta T_{>f} > 2·7$ °C and $\Delta T_{>f} > 4·0$ °C, respectively. (iii) Once the soil in the active layer was entirely thawed, θ_v was mainly affected by the infiltration of precipitation and showed no significant correlation with T_s.

The quantity of soil phase transitional water (Q) under different vegetation covers was estimated using Equation (3) (Tables VI, VII). At a soil depth of 1·20 m, the value of Q for alpine frost meadow soil under 30% cover averaged 15–55% and 10–47% (freezing phase) or 41–80% and 1·6–54% (thawing phase) more than such soils under 93% and 65% cover, respectively. Under both freezing and thawing processes, at a given soil depth, Q increased as vegetation cover decreased. Thus, Q at different depths in the active layer was strongly influenced by vegetation cover, with Q at any given depth significantly increasing as vegetation cover decreased. For the thawing phase in particular, Q for alpine frost meadow soils under different levels of vegetation cover was greatest at a depth of 0·40 m, corresponding to the portion of the active layer with the lowest ${}^{f}T_s$ (Tables VI, VII). This indicates that Q was related to ${}^{f}T_s$.

CONCLUSIONS AND DISCUSSION

θ_v and its dynamics in the active layer of a typical alpine frost meadow in the permafrost region of the

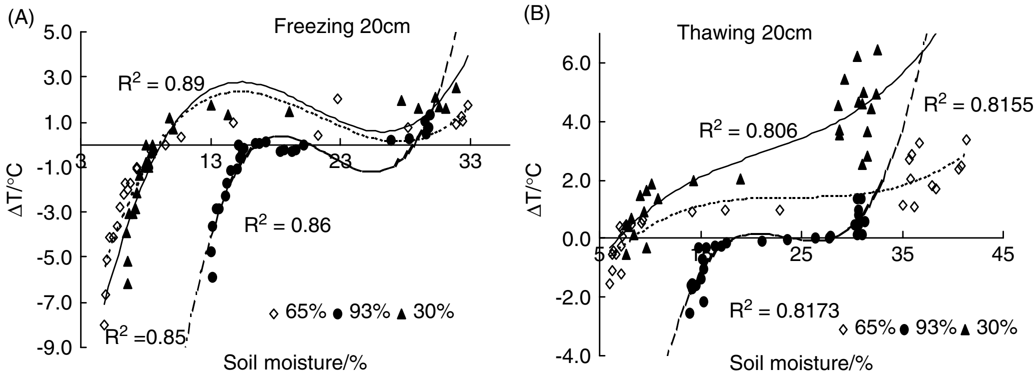

Figure 9. Soil water–soil temperature coupling and changes in profile under different vegetation covers of an alpine frost meadow grassland soil of the Qinghai-Tibet plateau

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Qinghai-Tibet plateau have an important influence upon water cycling and the growth of plateau meadow vegetation. Indeed these results indicate that vegetation cover has a significant impact on θ_v and its changes across the active layer of permafrost soils.

During both soil freezing and thawing periods, the lower the vegetation cover, the earlier are t_d and t_s, and the greater are ΔW_d, ΔW_s, and Q. While the active layer's t_z decreased as vegetation cover decreased, t_r decreased as vegetation cover decreased. The lower the vegetation cover, the lower the ${}^fT_s$ during the freezing period, but the higher the ${}^tT_s$ during the thawing period. During the thawing period, soil water tended to accumulate in the shallower and deeper portions of the active layer; as a result, θ_v in the middle (0·70 m) depth range was generally less than 25%, thereby forming a 'dry soil layer'. During the freezing period, this layer only occurred in alpine frost meadow soil under 93% vegetation cover. During the soil's entirely frozen period (a decrease in vegetation cover) the θ_v of the middle layer gradually increased, while θ_v in the lower layer (1·20 m) decreased.

When $\Delta T_{>f} < 5\,^\circ$C a significant cubic correlation existed between θ_v and T_s, regardless of the level of vegetation cover. As $\Delta T_{>f}$ exceeded 1·5 °C, the soil temperature at which the θ_v reached an approximate equilibrium point increased as vegetation cover decreased.

During the freezing period, precipitation in the study region was less than 25 mm (October–April), while evaporation exceeded 30 mm. The snow cover was irregular, filmy and discontinuously distributed over the ground surface, even in the middle of winter (Sato, 2001; Zhou *et al*., 2000). Therefore, the role of snow cover in the soil water-temperature coupling relationship and its changes under vegetation cover variations were not as those reported in other permafrost regions, such as those in North America and Siberia (Woo and Winter, 1993; Christensen *et al*., 2004), and was ignored in this study.

As a result of a decrease in vegetation cover, tied to degradation of alpine frost meadows on the Qinghai-Tibet plateau, soil water content changes in the active layer of such soils was exacerbated, and the response of soil water distribution to temperature changes was intensified. Given that several of China's most important rivers, including the Yangtze and Yellow, originate in the alpine frost meadow permafrost regions of the Qinghai-Tibet plateau, great attention should be paid to regional hydrological changes which could be engendered by severe degradation of alpine frost meadow ecosystems.

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