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Source: Arctic, Antarctic, and Alpine Research, 46(2) : 459-470

Published By: Institute of Arctic and Alpine Research (INSTAAR), University of Colorado

URL: https://doi.org/10.1657/1938-4246-46.2.459

Arctic, Antarctic, and Alpine Research, Vol. 46, No. 2, 2014, pp. 459–470

Soil Moisture Dynamics in a Mountainous Headwater Area in the Discontinuous Permafrost Zone of northern Mongolia

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Abstract

Soil moisture has widely been identified as a key factor for vegetation distribution in semi-arid areas. In the forest-steppe ecotone of the Khentii Mountains in northern Mongolia, soil moisture is directly controlled by exposition, slope, the presence or absence of permafrost, and vegetation cover. This study investigates the distribution of soil moisture and highlights the effects of a recent wildfire on this fragile ecosystem. Steep southerly exposed slopes are permafrost free and covered with steppe vegetation. Here, relatively warm and dry soils prevail, and high drying rates were observed following precipitation events during the summer period. The less inclined northerly exposed slopes are covered with taiga and feature relatively cold and wet soils overlying permafrost. Following a wildfire, the mean thickness of the organic surface layer drastically decreased from 0.15 m in the pristine taiga to 0.03 m in a heavily burned forest. As vegetation removal directly reduced evapotranspiration, soils in the burned forest were significantly the wettest and soil drying was less pronounced. Simultaneously, permafrost degradation was enhanced due to a significant increase in soil temperature. Thus, the conversion of forest areas to steppe after wildfires appears to be a long-term and possibly irreversible process during the ongoing climatic trend.

DOI: http://dx.doi.org/10.1657/1938-4246-46.2.459

Introduction

A significant increase in mean annual air temperature (MAAT) between 1.8 and 4.0°C at the end of the 21st century is predicted to occur at high latitudes as well as in major parts of Central Asia, including Mongolia (MARCC, 2009; IPCC, 2007). Already over the last 40 years, substantial increases in air temperature and in the frequency of extreme precipitation events have been observed in the Hövsgöl area in northern Mongolia (Namkhaijanstan, 2006; Nandintsetseg et al., 2007). For northern Mongolia, Batima et al. (2005) and Sato et al. (2007) reported a significant warming accompanied by a decrease in precipitation during the summer within the past 47 years leading to an increased abundance of droughts over the past decade (Davi et al., 2010). During this period, especially after the transition from a socialist to a market-oriented economy since the early 1990s, water demands frequently exceeded the water availability, as described in an exemplary way for the Kharaa catchment by Priess et al. (2011).

Although meteorological data are spatially and temporally rare, tree-ring-based hydroclimate reconstructions have shown a high variability of hydroclimatic conditions over the last three centuries in northeast Mongolia, with alternating drought and wet conditions and a relatively wet era during most of the 20th century (e.g., Davi et al., 2013; Pederson et al., 2013). However, the recent changes with the ongoing drying trend are of particular importance regarding the rapid growth of the mining, agricultural, and urban sectors (Priess et al., 2011; Karthe et al., 2013).

The Khentii Mountains in the northeast of Mongolia are situated in the discontinuous permafrost zone, which is one of the most sensitive areas to climate warming in the world (Gunin et al., 1999; Yoshikawa et al., 2003; Shur and Jorgenson, 2007). The distribution of permafrost in this transition belt between the boreal and dry mid-latitude climates of the northern hemisphere is strongly influenced by exposition and vegetation cover, which directly and indirectly influence the surface energy balance (Dulamsuren et al., 2005; Dulamsuren and Hauck, 2008). According to the classification presented by Shur and Jorgenson (2007), this region can be characterized as a climate-driven, ecosystem-protected permafrost area, where permafrost can persist in undisturbed late-successional ecosystems. Thus, a forest-steppe mosaic is prevalent, with southerly exposed steppe slopes that are permafrost free and experience seasonal frost. Here, vertical water percolation is assumed to dominate (Ishikawa et al., 2005), and soil moisture has been found to be generally low (Li et al., 2007a; Dulamsuren and Hauck, 2008; Liancourt et al., 2012). In contrast, northerly exposed slopes are covered with taiga forests, and soils are influenced by the presence of permafrost, which acts as an impermeable layer that restricts vertical water infiltration and permits lateral downward water movement beneath the surface (Ishikawa et al., 2005).

Regarding the forest-steppe mosaic, the predicted increase in MAAT has the following capabilities: (1) To extend unfavorable conditions for tree growth by the enlargement of steppe zones in presently forested areas. This is due to the fact that the future increase in potential evapotranspiration (ET) is expected to exceed the higher precipitation input in many places in Mongolia (Dulamsuren and Hauck, 2008; MARCC, 2009), resulting in a net loss of forests (Dulamsuren et al., 2010b) as tree growth decline and tree mortality are expected to increase (Liu et al., 2013). The forest conversion to grassland may further be exaggerated by herbivory small mammals and insects as well as the propagation of diseases (MARCC, 2009; Dulamsuren et al., 2008, 2010a). (2) To degrade permafrost and increase the thickness of the active layer (IPCC, 2012). This can durably affect local hydrology (Ishikawa et al., 2005), and water stress can occur more frequently. The active layer is defined as the soil above the permafrost table that

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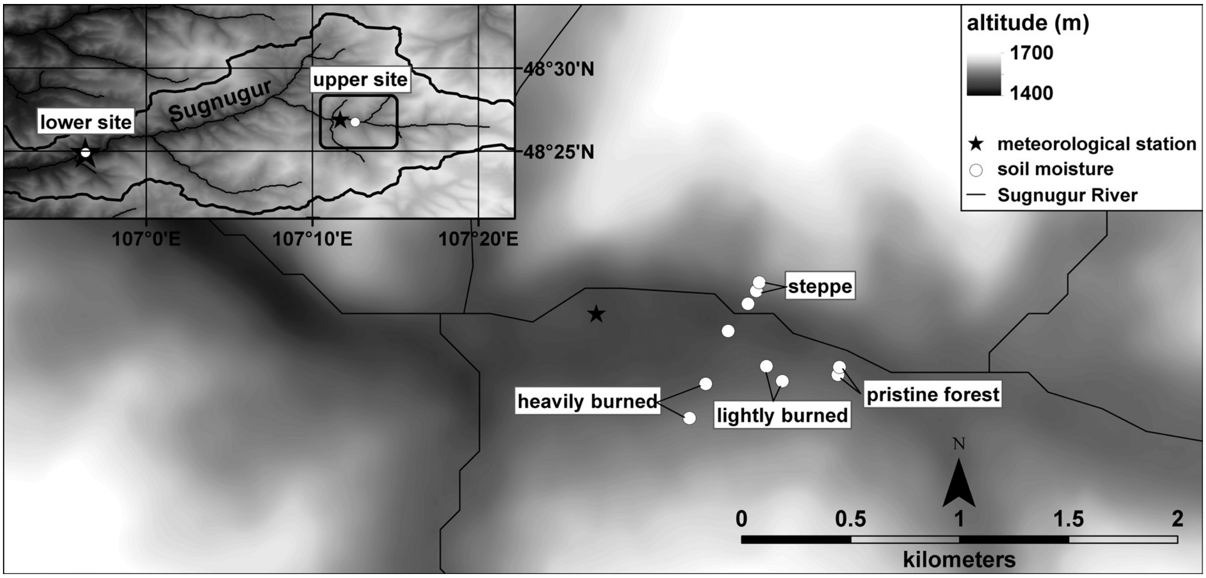

FIGURE 1. Topographic map of the Sugnugur basin, Mongolia. Shown are the positions of the lower study site (star, small map) and the upper study site (detailed map).

annually freezes and thaws. The depth of the frost table directly controls the rate of subsurface drainage during springtime as horizontal hydraulic conductivity of the saturated organic surface layer decreases exponentially with depth (Quinton et al., 2005). Degradation of permafrost alters the permeability and allows precipitation water to percolate deeper. This reduces soil moisture in the organic layer and the upper parts of the active layer (Ishikawa et al., 2005). (3) To shift the fire regime in boreal forests to a higher frequency of extreme fire years due to an augmentation of future water stress periods alongside increased human activity (Goldammer, 2002; Flannigan et al., 2009; MARCC, 2009; Tchebakova et al., 2009; Beck et al., 2011; Hessl et al., 2012; Saladyga et al., 2013). Depending on fire severity, wildfires can change hydro-meteorological processes, both immediately and in the long term, as they sustainably alter the ground thermal regime and the distribution of permafrost (Yoshikawa et al., 2003). Irreversible changes might occur if permafrost degrades or if combustion penetrates the insulating organic surface layer of the taiga (Hinzman et al., 1991; Harden et al., 2006). As the organic surface layer is reduced or removed, surface albedo decreases, soil thermal conductivity rises, the active layer thickness increases and permafrost degradation accelerates. Simultaneously, soil moisture in the upper layer decreases (Brown, 1983; Hinzman et al., 1991; Yoshikawa et al., 2003; Ishikawa et al., 2005). Regrowth of taiga following wildfires might, therefore, be hindered or possibly not occur at all. Burned forests may take up to 200 years or more to regenerate (Goldammer, 2002; Harden et al., 2006).

To our knowledge, few studies exist that deal with soil moisture distribution in a semi-arid forest-steppe ecotone (e.g., Dulamsuren and Hauck, 2008). A substantial number of studies examine the history and the effects of forest fires in regions with permafrost in Alaska (e.g., Yoshikawa et al., 2003; Harden et al., 2006; Bond-Lamberty et al., 2009; Flannigan et al., 2009; Barrett et al., 2011; Beck et al., 2011) and a few in Siberia and Mongolia (e.g., Tchebakova et al., 2009; Zhang et al., 2011; Lopez et al., 2012; Hessl et al., 2012; Saladyga et al., 2013). However, no study was carried out that highlights the effects of a recent forest fire on soil moisture distribution in the semi-arid and discontinuous permafrost zone of northern Mongolia.

The objective of the present study is to examine the soil moisture distribution and dynamics as well as the physical soil variables along a transect from a southerly exposed, permafrost-free steppe slope across a river floodplain to northerly exposed permafrost sites, which are covered by pristine, lightly and heavily burned forest. Our aim was to identify the most important controlling factors with regard to soil moisture distribution in a mountainous headwater area and to explore the effects of a wildfire on soil moisture conditions. We expect forest fires to cause an increase in solar radiation input, soil warming, and a short-term soil moisture increase and eventually long-term desiccation due to permafrost degradation.

Materials and Methods

STUDY SITE

Our study sites are located in the Sugnugur basin in the Khentii Mountains, approximately 100 km northwest of the Mongolian capital Ulan Bator (Fig. 1). The Khentii range is an alpine-type massif that stretches over the majority of northeastern Mongolia; it peaks at about 2800 m a.s.l. in the upper parts of the Sugnugur basin. The Sugnugur has been shown to contribute significantly to the discharge of the Kharaa (Menzel et al., 2011), the latter draining an area of approximately 15,000 km^2, including the western slopes of the Khentii and the dry steppe. Therefore, the Sugnugur plays an important role in the water supply of the rural population, which is mainly concentrated in the steppe lowlands.

Climate is highly continental and semi-arid with mean annual precipitation amounts ranging between 250 and 400 mm, of which 70% falls during the summer months (Menzel et al., 2011). The Sugnugur basin has a geological composition of granite and gneiss bedrock and is primarily sedimentary in the floodplain. The river course is mainly east-west directed; it thus subdivides the catchment into southerly and northerly exposed slopes, separated by the river's floodplain.

The lower study site, at an elevation of 1193 m a.s.l. (48°24′55″N, 106°56′21″E), is situated in the vicinity of a settlement where a small number of herders live (Fig. 1). The upper site in the headwaters, at an elevation of 1483 m a.s.l. (48°26′55″N, 107°11′41″E) (Figs. 1 and 2), is virtually unaffected by direct human impacts due to its remote location and poor accessibility. Although forest fires can frequently occur during dry periods in Mongolian boreal forests (Hessl et al., 2012), the study site was unaffected by forest fires since, at least, the beginning of the 1960s, as shown by satellite images (e.g., Corona and Landsat, data not shown). Due to a forest fire at the upper site in 2007, which destroyed considerable areas of forest stands with different intensities (lightly burned at the boundary areas, heavily burned at the center) on the northerly exposed slope, the effects of fire on soil moisture dynamics and ecosystem functioning could be investigated. Until today, the heavily burned area is very sparsely covered with vegetation as succession hardly proceeded.

At the upper study site (Figs. 1 and 2), the southerly exposed steppe slope is approximately 27.5° downslope and 31.5° upslope. It is characterized by herb-rich grasslands and dry, relatively warm soils without an organic surface layer. Shallow silty soils underlain by small boulders below a depth of 0.2 m prevail. The northerly exposed slopes exhibit cool and moist soil conditions and are covered by dense taiga consisting mainly of *Pinus sibirica, Larix sibirica,* and *Betula platyphylla* and a thick organic layer of mosses and small shrubs, predominantly *Ledum palustre* and *Vaccinium vitis-idaea*. Soils are dominantly silt-loam, which is gradually replaced by coarse gravel and small boulders below a depth of approximately 0.1 to 0.2 m. The slope is 11.5° to 12.0°, and the thickness of the organic layer above the mineral horizon, both on the downslope and upslope, averages 0.15 ± 0.04 m (n = 30). In contrast, there are major differences of the organic layer in the heavily burned forest. A dry, mostly dead leaf litter layer exists at the surface, with only a very sparse cover of vital mosses and shrubs, averaging a thickness of 0.05 ± 0.03 m (n = 30) downslope and 0.03 ± 0.02 m (n = 30) upslope. Thereby, the mineral horizon is partly exposed. The slope at the heavily burned forest ranges from 11.5° downslope to 14.5° upslope. The lightly burned forest has similar slope and soil characteristics as the pristine forest, and the thickness of the organic layer ranges from 0.07 ± 0.03 m (n = 20) downslope to 0.09 ± 0.02 m (n = 20) upslope.

INSTRUMENTATION AND MEASUREMENTS

A meteorological station at the lower site recorded (15 min intervals) air temperature, relative humidity (HMP155, Vaisala), and precipitation (1 min intervals) (Pluvio², Ott), among other parameters, with a CR3000 datalogger (Campbell Scientific). Before the main station was put into operation on 16 June 2011, a meteorological station (WXT5500, Driesen & Kern; weather transmitter WXT520, Vaisala) had recorded standard meteorological parameters since autumn 2010.

At the upper site, field measurements were carried out during the summer months of 2011 and 2012. A meteorological station (WXT5500, Driesen & Kern; weather transmitter WXT520, Vaisala) was installed in the floodplain from 25 June to 26 August 2011 and from 28 May to 3 September 2012. It measured standard meteorological parameters at 2 m above two-thirds of the vegetation height at 15 min intervals. Moreover, net radiation (1 min intervals) and soil temperature (15 min intervals) at a depth of 0.05, 0.1, 0.2, and 0.5 m were recorded. Due to datalogger failure, precipitation measurements are lacking from 14 July to 18 August 2012. During this period, precipitation recorded at the lower site was modified using altitude-dependent linear interpolation.

Volumetric soil water content (VWC) was measured at the upper site from 5 July 2011 until 3 September 2012 across three transects. Transect points were installed at the upper and lower slopes of the steppe, as well as in the heavily burned and the lightly burned forest. Additionally, VWC was measured at two sites in the floodplain. At each site, frequency domain reflectometry (FDR) sensors (10HS, Decagon Devices) were buried at depths of 0.05, 0.1, and 0.3 m in triple replicates to capture the influence of soil heterogeneity. A datalogger (EM 50, Decagon Devices) recorded VWC at 15 min intervals with an accuracy of ±3% using the standard 10HS sensor calibration equation. On 3 June 2012, the measuring points situated in the lightly burned forest were shifted toward the pristine forest, which allowed a better evaluation of the

FIGURE 2. View over the research area at the upper site. The steep left slopes are vegetated steppe, whereas slopes on the right are covered with burned and pristine taiga.

impact of forest fires. Over rainless days following precipitation events, soil drying rates were calculated as exponential decay (Liancourt et al., 2012). Four suitable periods could be identified in summer 2011 and seven periods in summer 2012. Soil hydraulic conductivity ($K_{f(unsat)}$) was determined on several occasions at each site using a minidisk infiltrometer (Decagon Devices) with different suction rates (Zhang, 1997). Suction rates of 2.0 and 0.5 cm were chosen to examine the $K_{f(unsat)}$ of the soil matrix and of the macropores, respectively.

Soil temperature was measured manually across the transects on 17 August 2011 to determine spatial differences. Boreholes were drilled with an Edelman driller (Eijkelkamp) in, at least, triple replicates. After drilling, soil temperature was immediately measured, using a Pt1000 sensor (Greisinger), at 5 cm intervals down to a maximal depth of 0.85 m, as far as boulders and frozen soils allowed. To assess spatial and temporal differences in temperature distribution across the transects, temperature sensors (DS1922L-F5, iButton) were installed in summer 2012 at the upper sites of the three transects, which continuously recorded soil temperature at hourly intervals from 23 June to 1 September. Three replicates at depths of 0.05, 0.1, and 0.3 m at each site, and additionally at a depth of 0.5 m at the heavily burned forest, accounted for soil heterogeneity.

LABORATORY ANALYSIS

To determine the water retention characteristics, undisturbed soil samples (100 cm³) were taken from the upper and lower steppe, lightly burned, and heavily burned sites at depths of 0.05, 0.1, and 0.3 m, directly adjacent (within 0.3 m) to the FDR probes, in June 2011. The analysis was performed using a pressure plate apparatus. VWC was measured gravimetrically during the drying phase at nine increments of pressure head, ranging from 0.01 m (near saturation [θ_s]) to 160 m (permanent wilting point [θ_r]). Water retention curves were fitted applying the van Genuchten (1980) analytical model.

STATISTICS

The aim of the statistical analysis using the statistics software R (http://www.r-project.org) was to ascertain the effects of the slopes with different vegetation types, inclination, exposure, and solar radiation on VWC, soil temperature, and infiltration capacity (both soil matrix and macropore infiltration) over the observation periods. To distinguish between interannual variations, VWC analysis was undertaken separately for the summer periods of 2011 and 2012. The normality of the data was assessed by the Shapiro-Wilk and the Kolmogorow-Smirnow tests, and a log-transformation was applied when required. Sphericity was verified by the Mauchly test and, if necessary, a Greenhouse-Geisser correction (Huynh-Feldt correction for small sample size) was made. A one-way analysis of variance (ANOVA) for repeated measures was applied to examine whether significant differences in the data sets exist. Afterwards, the paired *t*-test with Bonferroni correction ($\alpha = 0.05$) was used as a post hoc comparison to assess the pairwise differences between the slope sites.

Results

METEOROLOGICAL PARAMETER

During the year 2011 (2012), mean annual air temperature (MAAT) was -1.8°C (-3.0°C) and annual precipitation amounted to 317 mm (391 mm) at the lower site. The mean monthly air temperature varied between -28°C in January (2011 and 2012) and 16°C in August 2011 and July 2012 (Fig. 3). The precipitation amount of 220.4 mm during the summer months, June to August, 2011 was smaller compared to 275.2 mm in 2012 (Fig. 3). The upper site was generally 2 to 3°C colder during the summer measurement campaigns of 2011 and 2012 than the lower site. A daily mean air temperature of 12.4°C (2011) and 11.3°C (2012) was observed during the measurement period. Also summer 2011 was substantially drier with 171.4 mm of precipitation in comparison to 317.1 mm in 2012 (25 June to 26 August, respectively) and mean daily global radiation varied between -221.9 W m^{-2} (2011) and -195.9 W m^{-2} (2012). Fair-weather periods were characterized by daily air temperatures of around 20.0°C, daily relative humidity of 50%-60% and daily net radiation intensities of -190.0 W m^{-2}. These periods alternated with rainy days, with typical average daily air temperatures between 5.6°C (2011) and 2.9°C (2012), daily relative humidity of 90%, and daily net radiation intensities of around 0.0 W m^{-2}.

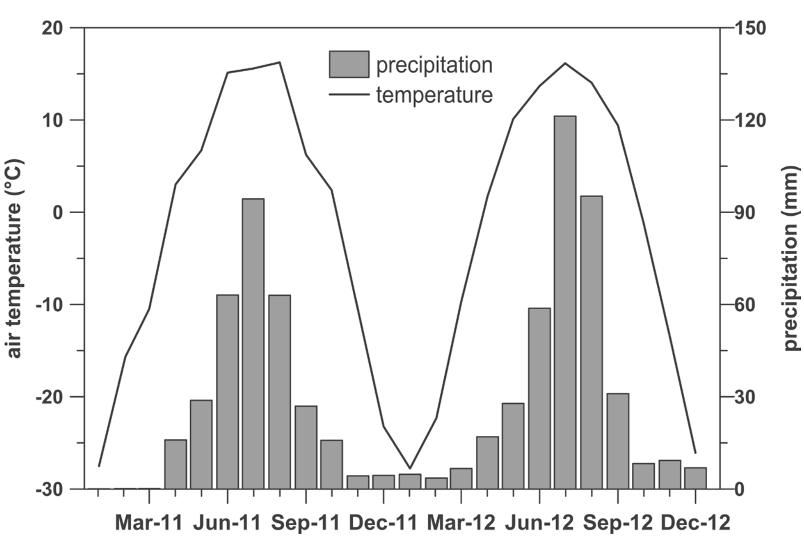

FIGURE 3. Monthly air temperature (°C) and precipitation (mm) values measured at the lower steppe site from January 2011 until December 2012.

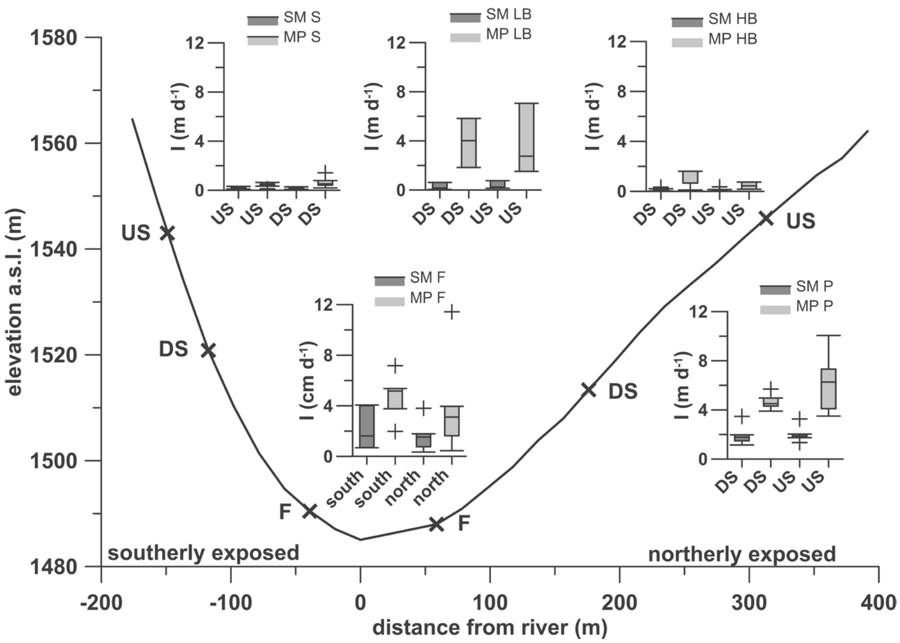

FIGURE 4. Measured unsaturated soil matrix (SM) and macropore (MP) infiltration rates (I) in m day^{-1} of the upslope (US) and downslope (DS) sites of the steppe (S), lightly burned forest (LB), heavily burned forest (HB), pristine forest (P), and the floodplain (F).

SURFACE INFILTRATION

A clear tendency was revealed by the distribution of the measured unsaturated hydraulic conductivity ($K_{f(unsat)}$) (n = 6) across the transects at the upper site (Fig. 4). Among all sites, the southerly exposed slope showed the lowest mean measured $K_{f(unsat)}$ values of the soil matrix (0.20 ± 0.08 m d^{-1}, up slope) and of the macropores (0.41 ± 0.18 m d^{-1}, up slope). At the northerly exposed slope, the highest values of $K_{f(unsat)}$ were observed at the pristine forest; mean rates occurred at the lightly burned forest; whereas the lowest rates were measured at the heavily burned forest. Thus, the pristine forest upslope matrix infiltration and macropore infiltration were 2.02 ± 0.65 m d^{-1} and 6.25 ± 2.36 m d^{-1}, respectively. In contrast, at the heavily burned forest upslope matrix infiltration of 0.13 ± 0.12 m d^{-1} and macropore infiltration of 0.40 ± 0.24 m d^{-1} were substantially lower. The observed mean upslope infiltration rates at the lightly burned forest of the soil matrix (n = 4) and of the macropores (n = 3) were 0.35 ± 0.29 m d^{-1} and 0.378 ± 0.29 m d^{-1}, respectively. $K_{f(unsat)}$ values at the downslope sites were similar across the transects, and high infiltration rates were measured at the river floodplain. The results of the ANOVA showed that soil matrix infiltration of the pristine forest was significantly higher compared to the highly burned forest (p = 0.02 upslope and p = 0.04 downslope) and steppe sites (p = 0.01 upslope and p = 0.04 downslope). Also, the macropore infiltration rates of the pristine forest were significantly higher compared to the heavily burned forest (p = 0.02 upslope and p = 0.00 downslope) and steppe sites (p = 0.03 upslope and p = 0.00 downslope).

TABLE 1

Mean soil temperature in °C and standard deviation at the up-slope sites during the observation period from 23 July to 1 September 2012, and minima and maxima of daily soil temperature values.

Site	Depth (m)	Mean soil temperature (°C)	Minimum soil temperature (°C)	Maximum soil temperature (°C)
Steppe	0.05	16.2 ± 0.5	9.98	21.22
	0.1	16.0 ± 0.3	10.75	20.41
	0.3	14.9 ± 0.3	11.82	17.78
Pristine forest	0.05	7.3 ± 1.5	2.71	11.61
	0.1	5.8 ± 1.6	1.63	9.02
	0.3	4.1 ± 1.6	0.94	6.45
Heavily burned forest	0.05	11.2 ± 1.8	7.17	15.01
	0.1	10.1 ± 2.4	6.87	13.21
	0.3	8.4 ± 2.6	5.63	10.53
	0.5	6.7 ± 2.7	4.07	8.27

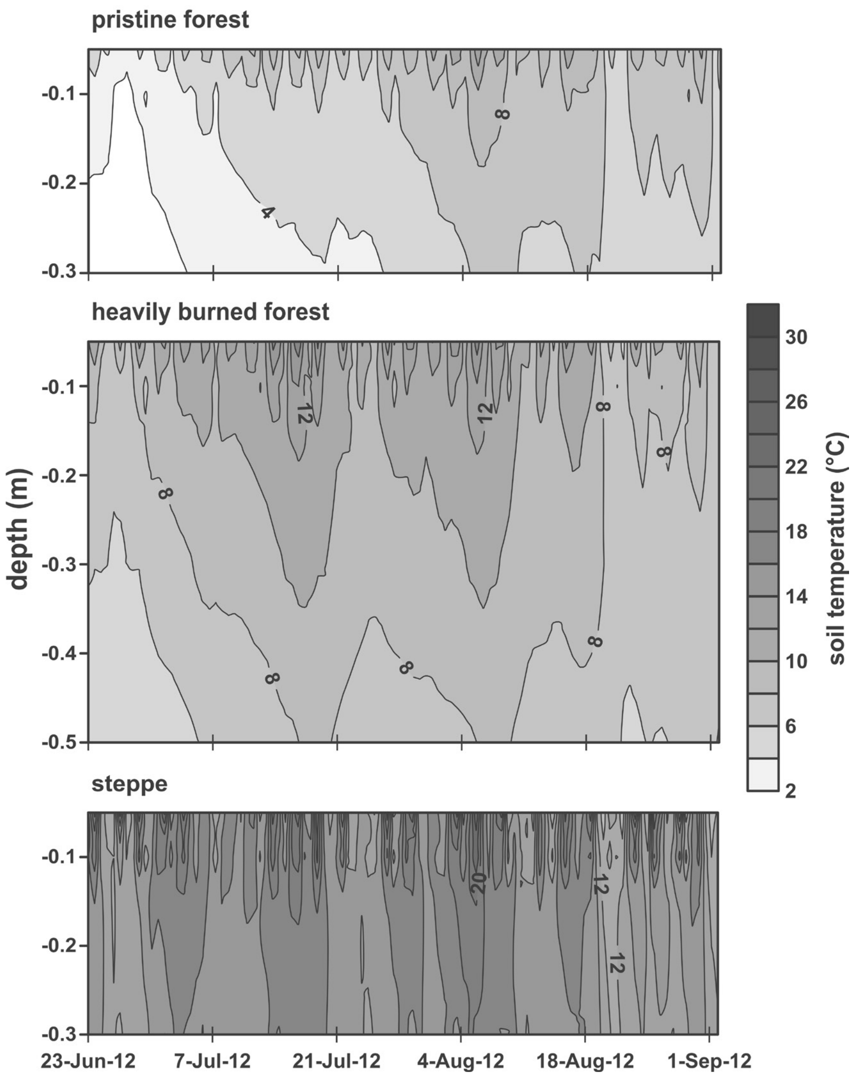

FIGURE 5. Mean soil temperatures (°C) at the upper slope of the pristine forest, the heavily burned forest, and the steppe site. Soil temperatures were measured in triple replicates at depth of 0.05, 0.1, and 0.3 m at all sites and additionally at 0.5 m at the heavily burned site.

SOIL TEMPERATURE, SOIL MOISTURE, AND SOIL DRYING

Soil temperatures showed characteristic, site-dependent behavior. During the manual measurements in mid-August 2011, the lowest soil temperatures were observed in the pristine forest. With a mean value of 12.8°C at a depth of 0.3 m, the southerly exposed site was the warmest, followed by the floodplain (11.3°C), the heavily burned forest (7.2°C), the lightly burned forest (4.5°C), and the pristine forest (2.7°C). Additionally, a northerly exposed terrace, vegetated only by grasses and shrubs at the slope bottom, was examined. Here, deeper drilling was possible and frozen soils were found at a depth of 0.7 m. During the summer months of 2012, temperatures were highest at the southerly exposed slope, where also the diurnal amplitude was most pronounced (Table 1, Fig. 5). The heavily burned forest was characterized by higher mean temperatures than the pristine forest (1.5 times, 1.7 times, and 2.0 times at depths of 0.05, 0.1, and 0.3 m). Results of the ANOVA revealed significant differences among all slopes (p values $< 2.0 \times 10^{-16}$).

At all sites, the mean winter volumetric water content (VWC) was small. During the summer periods of 2011 and 2012, the lowest mean values of VWC were found at the southerly exposed slope at all depths and the standard deviation was low. Also, the amplitude of the recorded VWC values at this slope was greater, and the response to rainfall events was most pronounced at depths of 0.05 and 0.1 m. Most strikingly, the mean VWCs during the comparison period from 5 July until 3 September were approximately 1.9 times higher at all depths in 2012 than during the drier summer of 2011. Interannual variability and short-term soil moisture dynamics at the northerly exposed slopes were less pronounced, and the mean VWC values were higher, with relatively constant high values at the heavily burned forest (Fig. 6). Comparing the soils of the pristine and heavily burned forest, it is evident that VWC at a depth of 0.05 m was drier in the pristine forest. The ANOVA results revealed that VWCs were significantly different among the slopes at each depth during the summer period 2011 and 2012 (p values $< 1.0 \times 10^{-3}$).

The determined mean soil drying rates across the transects are depicted in Figure 7. At a depth of 0.1 m, the rates were highest (23%) at the southerly exposed slopes in summer 2011 and slightly higher (15%) during summer 2012 compared to the pristine forest (11%). The smallest mean soil drying rates at a depth of 0.1 m were observed at the heavily burned forest with values of 4% (3%) during summer 2011 (2012). Generally,

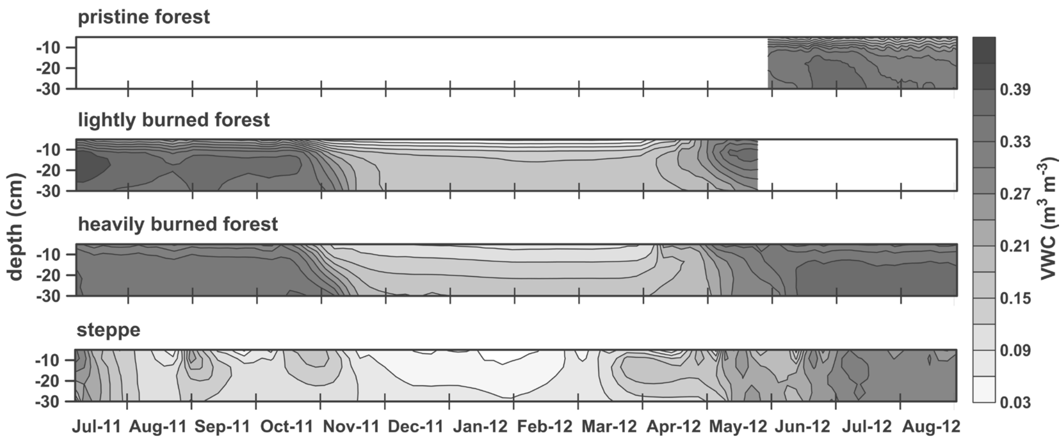

FIGURE 6. Volumetric soil moisture contents (VWC) in $m^3\ m^{-3}$ across the transects at depths of 0.05, 0.1, and 0.3 m from 5 July 2011 to 3 September 2012. Depicted are the mean values of the down- and up-slope plots at each site. FDR sensors of the lightly burned forest were removed at the beginning of June 2012 and installed in the pristine forest.

smaller rates were observed during the wetter summer 2012. The drying rates of the lightly burned forest were between the above-mentioned rates of the pristine and heavily burned forest during summer 2011.

PRESSURE PLATE ANALYSIS

The results of the pressure plate analysis indicate that the residual moisture content (θ_r) at the permanent wilting point (PWP) ranged between 0.07 and 0.11, 0.10 and 0.16, as well as 0.08 and 0.11 $m^3\ m^{-3}$ for the upper steppe, lightly burned, and heavily burned forest, respectively (Fig. 8). The mean VWC at the upper steppe slope during the dry summer period of 2011 was 0.15 $m^3\ m^{-3}$ (Table 2), which corresponds to soil tensions of about 7.6 m. However, minimum VWC at all depths dropped to values below the PWP (>160 m) during drought periods, resulting in heavily water-stressed conditions. In 2012, during the comparison period from 5 July to 3 September, mean VWCs were considerably higher (0.27 $m^3\ m^{-3}$) and mean soil tensions were around 1.8 m; minimum VWC values were clearly above the PWP (Table 2). However, VWC at a depth of 5 cm was close to the PWP at the end of May and mid-June (Fig. 6). Mean and minimum VWC values at both the lightly and heavily burned sites were above the PWP in 2011. Water-stressed conditions did not occur at these sites during the observation period.

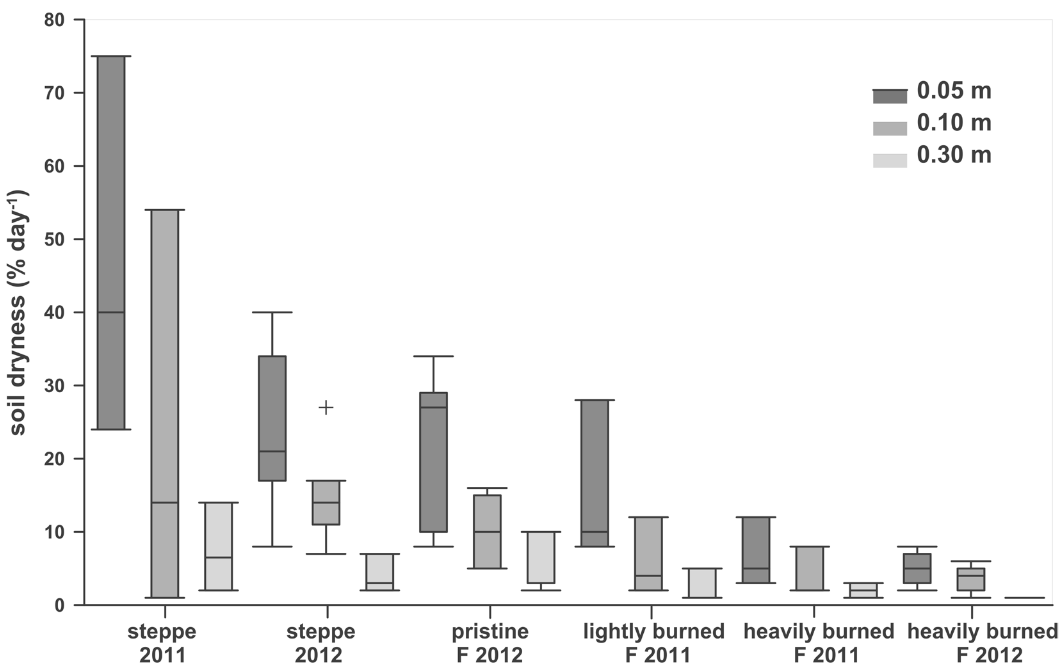

FIGURE 7. Determined soil dryness rates in $\%_v$ day^{-1} for the summer period of 2011 (n = 4) and 2012 (n = 7). The lightly burned sites were shifted toward the pristine forest at the beginning of June 2012.

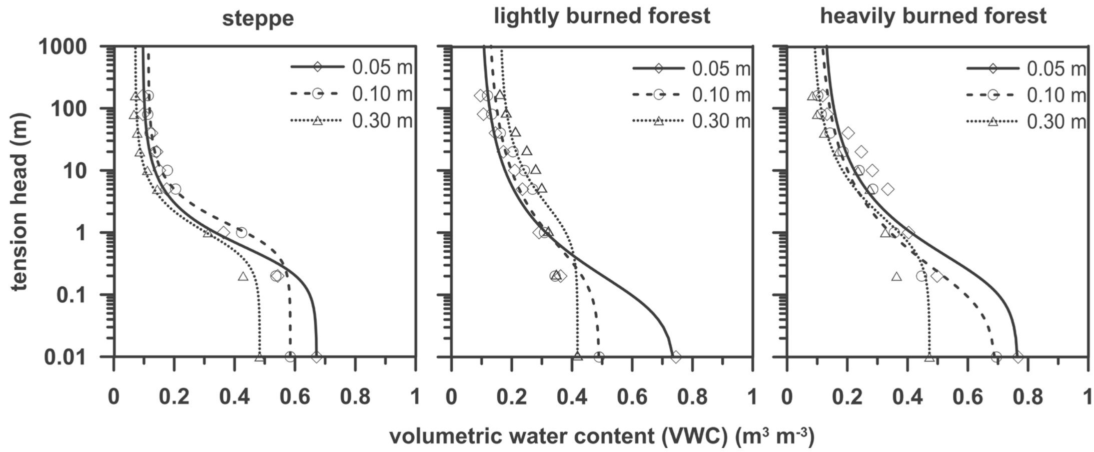

FIGURE 8. Results of the variation in soil moisture content determined from the pressure plate analysis of the soils from the steppe, lightly and heavily burned forest slopes, each at a depth of 0.05, 0.1, and 0.3 m. The best-fit curves were defined according to the method proposed by van Genuchten (1980).

Mean VWCs at the pristine forest and the heavily burned sites were 0.25 $m^3\ m^{-3}$ and 0.30 $m^3\ m^{-3}$ in 2012. As the pristine forest was not analyzed by the pressure plate method, we compared measured VWCs at a depth of 0.05 m with the water retention curve of the lightly burned forest. This analog could be conducted as soils of the pristine forest are comparable to those of the lightly burned forest with regard to the composition of the organic surface layer. As the mean VWC during the summer of 2012 at this depth was 0.15 $m^3\ m^{-3}$ (soil tension of 26 m) and even dropped to minimum values of 0.10 $m^3\ m^{-3}$ (pressure head >160 m), water-stressed conditions might have occurred. The results for the lower sites were similar across the transects (data not shown).

Discussion

Significant differences in the distribution of volumetric soil water content (VWC) in time and space across the three transects were observed (Fig. 6). During drought periods in summer, a clear tendency was revealed from warm and dry soils beneath the steppe vegetation toward cold and wet soils of the taiga forest (Figs. 5 and 6). This can primarily be derived from differences in exposition, slope inclination, plant coverage, and permafrost distribution (Ishikawa et al., 2005; Dulamsuren and Hauck, 2008; Liancourt et al., 2012).

STEPPE

At the permafrost-free, southerly exposed steppe slopes, intense net radiation resulted in a strong diurnal soil temperature pattern during the summer of 2012, as temperatures increased up to a maximum of 29°C at depth of 0.05 m in the afternoon and declined to a minimum of 7°C in the early morning. The high daily soil drying rates of 47% at a depth of 0.05 m (Fig. 7) and relatively low precipitation resulted in water-stressed periods in 2011, when the VWC temporarily dropped below the permanent

TABLE 2

Daily mean volumetric water content (VWC) in $m^3\ m^{-3}$ during the observation period from 5 July until 3 September, 2011 and 2012, across the soil moisture transects. Minimum and maximum recorded VWCs are listed to account for the soil moisture dynamics at each site.

Site	Depth (m)	Mean VWC ($m^3\ m^{-3}$) 2011 / 2012	Minimum VWC ($m^3\ m^{-3}$) 2011 / 2012	Maximum VWC ($m^3\ m^{-3}$) 2011 / 2012
Steppe	0.05	0.15 ± 0.03 / 0.27 ± 0.03	0.03 / 0.23	0.31 / 0.32
	0.1	0.16 ± 0.03 / 0.29 ± 0.02	0.07 / 0.25	0.32 / 0.33
	0.3	0.15 ± 0.03 / 0.29 ± 0.02	0.09 / 0.23	0.27 / 0.31
Heavily burned forest	0.05	0.29 ± 0.09 / 0.30 ± 0.06	0.26 / 0.28	0.31 / 0.31
	0.1	0.33 ± 0.10 / 0.34 ± 0.10	0.32 / 0.33	0.36 / 0.36
	0.3	0.35 ± 0.06 / 0.38 ± 0.03	0.34 / 0.38	0.36 / 0.39
Lightly burned forest (2011)	0.05	0.21 ± 0.10 / 0.15 ± 0.09	0.18 / 0.10	0.26 / 0.18
and pristine forest (2012)	0.1	0.34 ± 0.09 / 0.27 ± 0.09	0.30 / 0.23	0.40 / 0.32
	0.3	0.36 ± 0.05 / 0.35 ± 0.08	0.35 / 0.32	0.39 / 0.39

wilting point (PWP) (Fig. 6). This limitation in plant-available water during drought periods can explain the observation that, on the southerly exposed slopes, conifers are not able to grow and steppe vegetation dominates (Etzelmüller et al., 2006; Heggem et al., 2006; Dulamsuren et al., 2008). During the wetter summer period of 2012, the VWC did not drop below the PWP and daily soil drying rates were lower (24% at a depth of 0.05 m) due to higher VWC values and precipitation amounts compared to 2011 (Figs. 6 and 7). A similar observed difference in soil drying rates between two years, with smaller values during the wetter year, was reported by Liancourt et al. (2012) for a southerly exposed steppe slope in northern Mongolia. In summer 2012, daily actual evapotranspiration (ET) rates peaked at around 6 mm d^{-1} after periods with high precipitation input and thus relative high VWC, whereas actual ET rates gradually declined (<1 mm d^{-1}) during water-stressed conditions, as examined by Minderlein (unpublished data) using the Bowen ratio energy balance method.

Steep slopes and low infiltration rates (Fig. 4) reduced the amount and depth of percolating precipitation water. Occasionally, overland flow could be observed during high-intensity precipitation events. The observed low VWCs of <0.1 m^3 m^{-3} at the southerly exposed slopes during drought periods are in accordance with the studies of Li et al. (2007a), Dulamsuren and Hauck (2008), and Liancourt et al. (2012). However, high daily VWC values of >0.3 m^3 m^{-3}, as observed in our study at the southerly exposed steppe slopes during rainy periods, were not reported earlier.

PRISTINE FOREST

The pristine forest at the less inclined northerly exposed slope exhibited similar topsoil drying rates as the steppe slope during the growing season of June, July, and August 2012 (Fig. 7). However, the VWC remained high at a depth of 0.3 m, with only small variations following precipitation events (Fig. 6). This can be attributed to the storage of a considerable amount of water in the active layer, which is underlain by the impermeable permafrost (Sugimoto et al., 2002; Ishikawa et al., 2005; Etzelmüller et al., 2006). As soil temperature at a depth of 0.3 m gradually increased from 0.9°C at the end of June to a maximum of 6.5°C at the beginning of August 2012, the underlying permafrost degraded deeper into the mineral horizon and the thickness of the active layer increased (Ishikawa et al., 2005; Quinton et al., 2005; Heggem et al., 2006). Sugimoto et al. (2003) observed a maximum thickness of the active layer of 1.2 to 1.5 m at the end of August underneath a larch forest in East Siberia near Yakutsk. Precipitation water, infiltrating at significantly highest rates (Fig. 4) into the organic surface layer, now percolated deeper into the soils. The observed decrease in VWC at a depth of 0.3 m from mid-July after the soil thawed is coherent with results reported by Takata (2002) and Sugimoto et al. (2003). Using a one-dimensional land surface model, Takata (2002) showed that a wet soil zone is formed above the thawing front. This zone moves down as thawing progresses and soils in the upper part are drained. Subsequently, the previous shallow lateral subsurface drainage in the organic layer with high hydraulic conductivity is stopped (Quinton et al., 2005), except for excess water during periods with high precipitation input.

Mean daily ET rates of the boreal canopy are reported to account for approximately 1.4 to 2.3 mm d^{-1} (Ohta et al., 2001; Lopez et al., 2008; Ohta et al., 2008). The organic surface layer in the pristine forest reduces VWC at the surface during drought periods due to daily ET rates ranging between 0.3 and 1.5 mm d^{-1} (Heijmans et al., 2004; Bond-Lamberty et al., 2009). This considerable loss resulted in the relatively high soil drying rates and low VWC at the soil surface at a depth of 0.05 and 0.1 m observed for the pristine forest in our study. As water-stressed conditions might occur during drought periods at a depth of 0.05 m, the spreading of forest fires may be facilitated.

Thawing permafrost plays an important role as a direct source of water for the larch trees during drought periods, as observed by Sugimoto et al. (2002). Li et al. (2007b) found soil water in the uppermost 0.3 m to be the main source for larch trees in months with little precipitation input. In return, the taiga forest protects the permafrost due to reduced net radiation input through canopy interception and by the thermal insulating capacity of the organic layer (Brouchkov et al., 2005; Harden et al., 2006; Heggem et al., 2006; Zhang et al., 2011). Yoshikawa et al. (2003) stated that an organic surface layer with a thickness of 0.1 m provides adequate thermal resistance to protect the frozen mineral soil.

BURNED FOREST

Following deforestation, shadowing, organic layer thickness, and albedo are decreased, yielding a greater absorption of shortwave radiation at the ground's surface. Simultaneously, soil properties such as soil density and moisture, infiltration and evaporation rates, thermal conductivity, and heat capacity are altered (Hinzman et al., 1991; Yoshikawa et al., 2003; Brouchkov et al., 2005; Dulamsuren et al., 2005; Iwahana et al., 2005; Harden et al., 2006; Etzelmüller et al., 2006). We observed a decrease in vital organic surface cover in the lightly burned area and a removal in most areas of the heavily burned forest. Yoshikawa et al. (2003) reported that the ratio of ground heat flux and sensible heat flux to net radiation increased following fire, whereas the latent heat flux decreased. The authors stated that differences in temperature between soils of burned and unburned forests have been shown to correlate with the thickness of the organic and active layer. Therefore, thermal conductivity increases after the organic layer is reduced or removed. This effect could be observed at the heavily burned forest as soil temperatures were significantly higher compared to the pristine forest at all depths ($p < 2.0 \times 10^{-16}$). This is enforced by the prevalent wet soils at these sites as high VWCs are reported to further increase thermal conductivity (Takata, 2002; Harden et al., 2006; Shur and Jorgenson, 2007). Depending on fire severity, the above-mentioned factors can explain the differences in soil temperature between our northerly exposed sites for the years 2011 and 2012 (Fig. 5).

As a consequence of the reduced ET after the removal of the vegetation, significantly highest mean VWCs across the transects were found at the lightly and heavily burned forest in 2011 and 2012, respectively (Table 2, Fig. 6). Hence, as transpiration came to a virtual standstill, calculated soil drying rates were smallest at the heavily burned sites (Fig. 7). This finding is in accordance with Iwahana et al. (2005), who observed soil water retention after one year at a cutover site compared to a reference forest site near Yakutsk, eastern Siberia. Yoshikawa et al. (2003), who studied postfire behavior of boreal forests in Alaska, reported short-term soil moisture increases after wildfires as transpiration rates decreased due to the loss of vegetation and long-term (more than a decade) drying due to an increasing thickness of the active layer and the recovery of vegetation. The increased soil temperatures of the heavily burned forest clearly indicate enhanced permafrost degradation to deeper depths as we recorded temperature values up to 8°C at a depth of 0.5 m in July and August (Fig. 5). Iwahana et al. (2005) stated that vegetation removal enhanced ground thawing and therefore increased the thickness of the active layer by 0.14 m after one year.

These changes in active layer thickness affect local hydrology. Infiltrating precipitation water now percolates deeper into the warm and more permeable permafrost (Brouchkov et al., 2005; Ishikawa et al., 2005; Quinton et al., 2005), possibly resulting in thermokarst (Iwahana et al., 2005) and drier soils in the long term (Yoshikawa et al., 2003). This may further exacerbate permafrost recovery, as the conduction of cold winter temperatures toward the underlying soils is reduced in dry soils (Harden et al., 2006). Typically, a time span occurs of 3 to 5 years after boreal forest fires during which the active layer increases to a thickness that does not completely refreeze the following winter (Brown, 1983). DeBano (2000) reported a near-surface hydrophobic layer after burning that can resist surface water infiltration, which is coherent with the significantly lower surface infiltration rates measured in the heavily burned forest compared to the pristine forest (Fig. 4). This reduced infiltration capacity might further limit recharge in the active layer following precipitation events and increase hillslope runoff, as the severity of water repellency at lower soil water contents is enhanced, especially during the dry season (DeBano, 2000).

Conclusions

This study examined the distribution of different soil parameters in a semi-arid forest-steppe ecotone. In particular, the effects of a recent forest fire (2007), which was characterized by different burn intensities, in the Khentii Mountains in northern Mongolia were investigated. The hydroclimatic conditions in this region have been shown to undergo frequent changes over the past three centuries (Pederson et al., 2013). However, the current climatic conditions can sustainably alter this fragile ecosystem, which can be considered to be representative of other semi-arid mountainous areas in northern Mongolia. Considering the increasing anthropogenic and natural disturbance regime, as for example by wild food gatherers, increasing MAAT, and drought conditions, these are likely to result in declined tree growth and increasing tree mortality (Dulamsuren et al., 2010b; Liu et al., 2013), as well as in a propagation of forest fires (Hessl et al., 2012). Therefore, the forest cover and the distribution of the coupled permafrost-forest system are decreasing. Once permafrost is degraded, water stress generally increases on the long-term, resulting in an expansion of steppe areas. Although our results show an increase in soil moisture 4 and 5 years after a forest fire, the study of Yoshikawa et al. (2003) gave a strong hint toward long-term drying and propagation of steppe vegetation (Dulamsuren et al., 2010b) that is adapted to water limitation during drought periods.

As mountain forests form the main water-generating areas in the forest-steppe ecotone (Karthe et al., 2013), their future development must be of special interest, especially considering the growing demand for water in urban, agriculture, and mining development (Priess et al., 2011; Karthe et al., 2013; Pederson et al., 2013). Hence, longer observational studies of soil moisture distribution and of the rate of vegetation change following forest fires in the forest-steppe ecotone of northern Mongolia are necessary.

Acknowledgments

The authors thank Steffen Füssel for help during field work and Gerd Schukraft for help in the laboratory. Without a variety of Mongolian colleagues, namely Chimegsaikhan Altangerel, Nasanbayar Narantsogt, Khuhuu Gunsenbat, and Nugonhayaar Gansukh, just to name a few, this study could not have been realized successfully. We also acknowledge comments of two anonymous reviewers, which greatly helped to improve this paper.

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MS accepted November 2013