EARTH'S CRYOSPHERE
SCIENTIFIC JOURNAL

Kriosfera Zemli, 2019, vol. XXIII, No. 4, pp. 3–12 **http://www.izdatgeo.ru**

REGIONAL AND HISTORICAL GEOCRYOLOGY

DOI: 10.21782/EC2541-9994-2019-4(3-12)

SEASONALLY FROZEN LAYER OF PEATLANDS IN THE SOUTHERN TAIGA ZONE OF WESTERN SIBERIA

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This paper presents results of the 2011–2017 field studies of the thickness of seasonally frozen layer and the temperature regime of peat soils of oligotrophic bogs in the southern taiga zone of Western Siberia. The experimental observations have been carried out using the atmosphere-soil measurement system at different depths from the surface down to 240 cm. The five study sites (plots) included: a hollow and ridge at the ridge-hollow bog complex, open fen, high ryam and low ryam. The analysis result has revealed significant differences in the temperature regime of peat soils and the depth of seasonal freezing at plots with high and low levels of bog waters. The proposed regression model allows to predict the maximum depth of seasonal freezing, using the data on maximum snowpack height and winter mean air temperature.

Peatland ecosystems, microclimate, soil temperature regime, seasonally frozen layer, freezing depth, snowpack, Western Siberia

INTRODUCTION

Peatlands, the unique natural wetland landscapes participating in the regulation of atmospheric chemical composition, play an integral role in global water balance of the biosphere and biological diversity [*Dokturovsky, 1932; Boch and Mazing, 1979; Liss et al., 2001; Barid et al., 2013*]. Boreal peatland ecosystems, in themselves, are long-term carbon sequestering systems (aka "carbon sinks") [*Ivanov and Novikov, 1976; Vomperskii et al., 2005; Rydin and Jeglum, 2015*] containing according to various estimates 120–455 bln ton of carbon [*Efremov et al., 1994; Kudeyarov et al., 2007; Bukvareva, 2010; IPCC, 2013*], of which more than 70 bln ton are held in the West Siberian peatlands [*Sheng et al., 2004*]. As such, the appreciable amounts of released carbon when affected by climate change or anthropogenic impact lead to its rising fraction in the atmosphere (as CO_2 or CH_4) and significantly contribute to the global carbon cycle [*Semenov, 2004; Karelin and Zamolodchikov, 2008; Anisimov et al., 2012; Desyatkin et al., 2015*].

Effects of climate warming [*IPCC, 2013; The Second... Report..., 2014; Anisimov and Kokorev, 2017*] generally entail variations in thermal regime of soils (both in peat and mineral soils) and frost penetration depth [*Vomperskii, 1968; Pavlov, 2008; Sherstyukov and Sherstyukov, 2015; Desyatkin and Desyatkin, 2017; Peng et al., 2017*], as well as in the timing of snowpack deposition and degradation [*Osokin and Sosnovsky, 2014; Dyukarev, 2015; Wang et al., 2015; Zhong et al., 2018*], and influence the snowpack accumulation processes [*Kitaev and Kislov, 2008; Pavlov, 2008; Osokin and Sosnovsky, 2015; Voropay and Vlasov, 2017; Wegmann et al., 2017*].

Soil temperature is a key factor controlling many biotic and abiotic processes in soils involving soil organic matter decomposition and mineralization, greenhouse gas emissions, and release of dissolved organic carbon [*Vomperskii, 1968; Arkhangelskaya, 2012; Golovatskaya and Dyukarev, 2012; D'Angelo et al., 2016*].

Thermal regimes of peat (i.e. organic) soil and mineral soil are known to differ essentially [*Chechkin, 1970; Ivanov and Novikov, 1976; Gilichinsky, 1986; Ershov, 1989; Ospennikov, 2001; Dyukarev et al., 2009; Trofimova and Balybina, 2015*]. Some fundamental characteristics of organic soil in peat deposit, which in itself is complex organo-mineral system, include: high porosity and, accordingly, saturation with water (due to its high hydraulic conductivity and weak suction), as well as large amounts of underdecomposed organic matter [*Romanov, 1961; Chechkin, 1970*].

The formation of seasonally frozen soil layer (active layer) is primarily driven by the incoming solar radiation and heat balance at the surface [*Kudryavtsev, 1981, p. 82–110; Ershov, 2004, p. 248–303*]. The major environmental controls of the earth-atmosphere energy balance determining the freezing and thawing processes in soils are: land cover type, topography, soil composition and moisture content, hydro-

logic conditions, etc. Peatlands respond to climate cooling by gradually transforming into permafrost peatlands with affiliated formation of frost heave mounds [*Kudryavtsev, 1981, p. 111–132*]. Beyond the area of permafrost distribution, the active layer becomes an independent natural phenomenon, controlled by the long-term, deep cryotic process. In this respect, the seasonally freezing upper layer of the lithosphere is seen as the object of both permafrost study and soil science [*Gilichinsky, 1986*].

Given that the West Siberian peatlands are characterized by relatively shallow soil freezing, the frost penetration depth, when increasing, is accompanied by essentially longer delay in the timing of complete annual thaw of soils, as compared to mineral soils [*Ivanov and Novikov, 1976; Chigir, 1978*]. The deepest frost penetration into soils (up to 2.5–3.0 m) is characteristic of sands [*Ershov, 1989*]. In forested areas, clay loams generally freeze to a greater depth (1.0–1.8 m), as compared to peat (0.3–0.9 m), which, if water-saturated, fails to freeze completely during the winter at many snow-covered sites [*Ershov, 1989*]. The pioneering studies of the freezing regime of peatlands in the Baraba lowland [*Serebryanskaya, 1946*] showed that the snowpack height (*H*) and frost penetration depth (*FD*) differentiated by the mesorelief heterogeneity (i.e. surface elevation), varies from 50–70 cm (raised bogs) to 85–160 cm (mineral ridges). Low-level blanket bogs (*FD* = 50–90 cm) and fore-peatland zone (*FD* = 85–110 cm) occupy intermediate position. The maximum frost penetration depth within the extent of West Siberian peatlands has increased from 76 cm (beneath sedge-hypnum hollows) to 102 cm (beneath sedge-hypnum complexes) [*Checkin, 1970*]. In northern European Russia, the thickness of the frozen layer of undrained peatlands differs from 24 cm (eutrophic and oligotrophic pine-sphagnum peatlands) to 62 cm (hummocky bogs of the northern Kola province) [*Chechkin, 1970*].

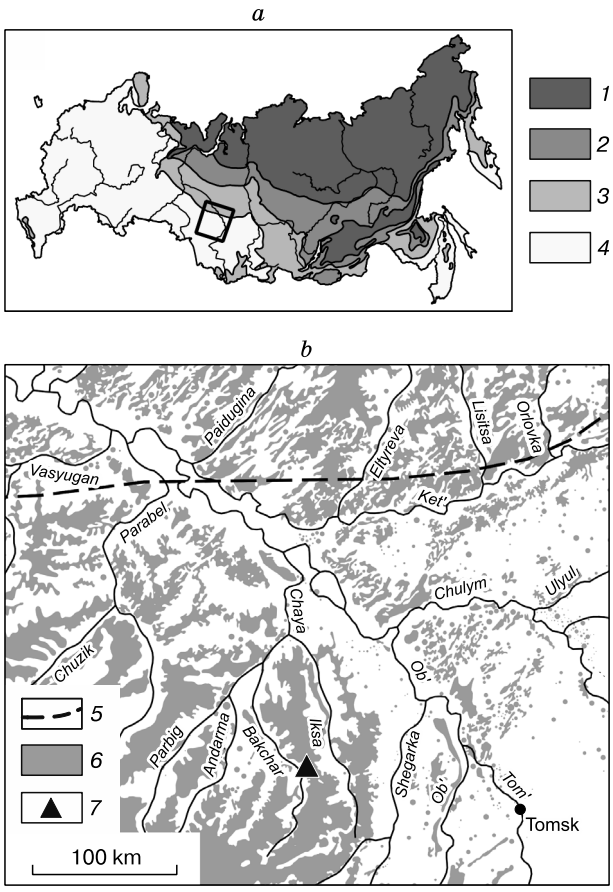

Fig. 1. Schematic layout of the study area.

a – distribution of permafrost (after [*Stolbovoi and McCallum, 2002*]): *1* – continuous, *2* – discontinuous, *3* – sporadic; *4* – seasonally frozen ground; *b* – the area of study: *5* – the limit of sporadic permafrost; *6* – peatlands [*Sheng et al., 2004*]; *7* – study area.

In the period spanning 1966–2012, the snowpack height showed a generally increasing trend during the winter and spring months across Northern Eurasia, whereas it was tending to decrease during the fall [*Zhong et al., 2018*], which has affected the active layer dynamics. Thus in response to warming surface air temperature (SAT) in the observation period (from 1978 to 2012) in the northern part of European Russia, the mean annual soil temperatures of peat deposit increased at a rate of 0.2–0.9 °C/10 years to a depth up to 320 cm due to the propagation of heat downwards into the peat column [*Kalyuzhny and Batuev, 2015*].

The study of microclimate characteristics of peatland ecosystems is critical in conducting comprehensive research [*Kabanov, 2015*]. Mean annual soil temperatures within peatlands can be both higher and lower, than in the surrounding non-waterlogged environments [*Ospennikov, 2001*], inasmuch as they are largely governed by the genesis and stage of bog evolution. Evaluation of the direct impacts of wetlands on regional climate will allow to correctly interpret the existing climate change scenarios for the next century. While the established patterns of peatland ecosystems functioning under different climatic conditions and analysis of their changes provide the basis for a reliable estimation of vegetation and multidirectional carbon fluxes. The results obtained are equally important for assessment of the contribution from the vast West Siberian wetlands to regional and global climate variations, rather than the global carbon cycle alone.

OBJECTS OF STUDY AND RESEARCH METHODS

The Vasyuganie geophysical field station of the Institute of Monitoring of Climatic and Ecological Systems of the Siberian Branch of the Russian Academy of Sciences (IMCES SB RAS) [*Golovatskaya et al., 2008*] in the Bakchar bog area was chosen to be the study area. Even though permafrost is reported as absent in this area (Fig. 1), a degrading frost heave

mound dated 300 to 400 yrs BP discovered at the bog massif periphery in 2007 serves as an indication of the southernmost limit of permafrost [*Dyukarev and Pologova, 2007*]. The study area can therefore be ranked as the zone of area-specific sporadic permafrost distribution [*Vasil'chuk, 2013*].

The five observation sites (plots) set up on the Bakchar bog and representative of major regional oligotrophic bog ecosystems [*Dyukarev et al., 2011*] are differentiated by their biogeocenoses as follows: pine-shrub-sphagnum (high ryam); pine-shrub-sphagnum with oppressed tree stand (low ryam); sedge-sphagnum fen; pine-shrub-sphagnum ridge; and scheuchzeria-sphagnum hollow within the ridge-hollow complex (RHC). The peat deposit thickness variation within the observation sites is characterized as: the highest (310 cm) in the fen and hollow within RHC; slightly decreasing on the ridge of RHC (230 cm) and in the low ryam (210 cm); not more than 100 cm in the high ryam. The peat deposit is underlain by clays, acting as an impermeable bed for bog waters.

The plots were divided into two main groups depending on the bog water table levels (WTL): (1) high WTL (5–10 cm above the surface) in fen and hollow of RHC; (2) low WTL (25–45 cm above the surface) in the plots comprising high ryam, low ryam, and ridge of RHC.

Soil and air temperatures were measured from April 1, 2011 through October 3, 2017 using atmosphere-soil measuring complex (ASMC) with 1 h measurement frequency [*Kurakov, 2012; Kiselev et al., 2017*]. Peat temperature was measured on the surface and at depths of 2, 5, 10, 15, 20, 30, 40, 60, 80, 120, 160, 240 cm. The mean diurnal and monthly temperatures, as well as the soil freezing depths were calculated from the initial data sets.

Soil freezing depth was calculated as the depth of 0 °C temperature penetration into soil [*Drozdov, 1957*]. The freezing depth was calculated by linear interpolation of spot soil temperature measurements between two adjacent depths, provided that one of them had a negative temperature. This was followed by calculation of the mean daily freezing depths, which is discussed below.

The data on the snowpack height were obtained from the nearest Bakchar weather station with referencing the RIHMI–WDC website [*Bulygina, 2017*]. The estimated snow cover parameters included: the snowpack height; the date of permanent snow cover formation and destruction; overwinter snowpack height variation and the date of its achieving the maximum value [*Drozdov, 1957*].

CLIMATIC CHARACTERISTICS OF STUDY AREA

The data reported from the nearest, Bakchar weather station located 30 km to the west of the study area included the mean annual air temperature (MAAT) for 1936–2017 (–0.3 °C) and mean temperatures of the warmest (July) and coldest (January) months (+18.1 °C and –19.2 °C, respectively). The mean monthly air temperature remained in the negative domain from November (–9.9 °C) through March (–8.9 °C); the lowest MAAT (–2.7 °C) was recorded in 1969.

The rate of MAAT increase is found to be 0.22 °C/10 years. The revealed temperature increasing trends were most significant for November (0.35 °C/10 years), December (0.49 °C/10 years), March (0.64 °C/10 years), February and April (0.29 °C/10 years).The trends in MAAT variations have been most remarkable over the last three decades [*Dyukarev, 2015*]. The sum of annual precipitation totaled 468 mm, of which 45 % falls in the summer months and 12 % during the winter season.

Arrival of the first ground covering snow cover (on average, October 13, for the period of 1936–2017) usually ends by its melting due to the subsequent thaw days. It's only after temperatures become steadily negative (on average, since October, 30) that permanent snowpack is settled for the winter. Its degradation is usually completed by April 20, with its duration averaging 172 days. The maximum snowpack height (23–113 cm) during the winter is observed in mid-March.

PEAT SOILS TEMPERATURE

The visual representation of difference in temperature conditions in the plots with high and low WTL is provided by analysis of the annual course of peat soil temperature at depths up to 240 cm. During the warm period, when air temperature values are close, while density and moisture content of peat differ slightly, the plots with high WTL tend to be heating better than those with low WTL. The maximum surface temperature within the RHC was observed in July, with the values varying from 18.1 °C (hollow) and 17.4 °C (fen) (Fig. 2), whereas in the plots with low ryam, high ryam and on the ridge the temperature values were: 17.3, 14.1 and 17.7 °C, respectively.

Besides the bog water table level, specific features of the temperature regime of peatlands during the warm season are affected by the vegetation cover as well [*Ershov, 2004, p. 248–303*]. Thus the 0–60 cm layer of the low and high ryam plots belonging to conventionally cool wetland ecosystems showed a lower temperature, than the same depth interval in the ridge of RHC. Given that the ridge is lacking dense vegetation, its surface therefore receives more insolation. Thus, during the warm season, the ridge appears the warmest bog ecosystem among the low BWL group. Alternatively, the low and high ryam plots receive less solar radiation which is partly absorbed by the dense woody vegetation, and, com-

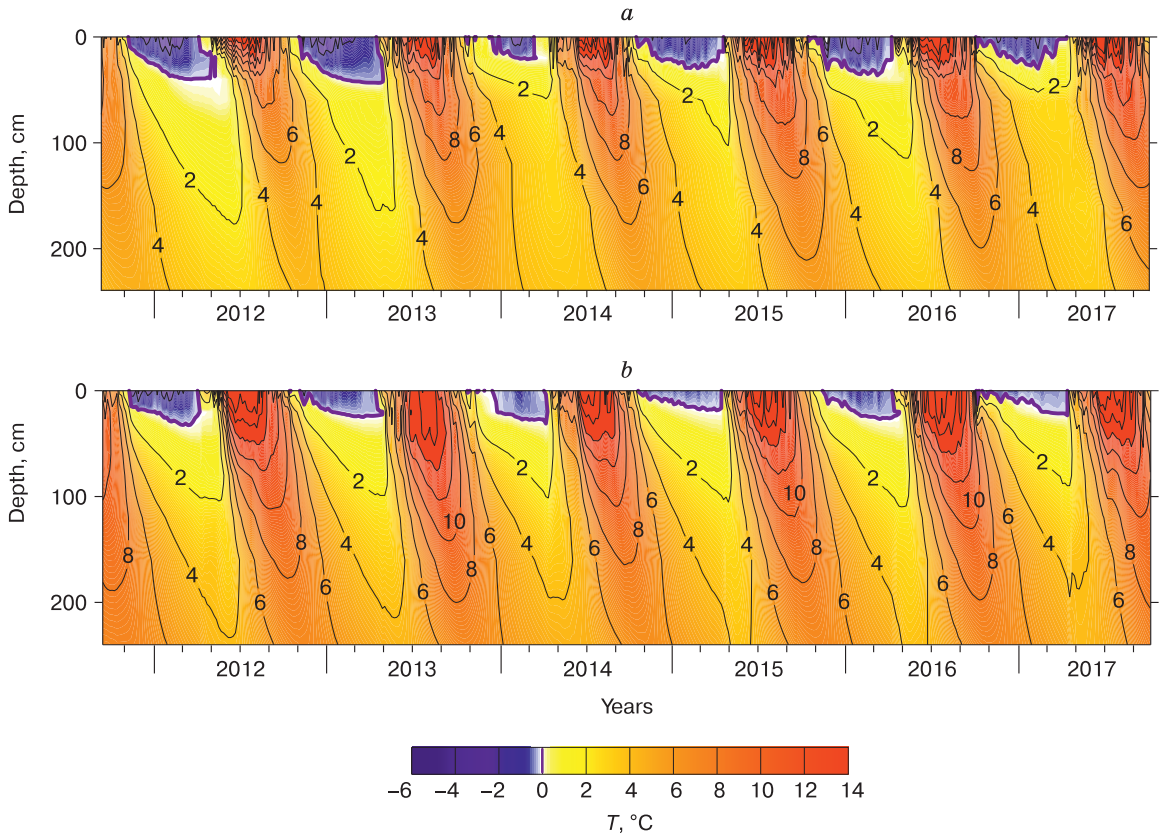

Fig. 2. Temperature of peat soil (*T*) measured in the plots: low ryam (*a*) and hollow of RHC (*b*).

plexed with the low-level water table, eventually become colder in comparison with fens, hollows and ridges of the RHC.

The reasons why the ridge of RHC is interpreted as the coldest area during the winter season are: (1) the snowpack height in the ryam plots is slightly higher due to snow-retention capacity of shrubs and tree trunks [*Evseeva et al., 2016*]; (2) peat layer is seen to be warmer in the hollows and open fens, even at shallower snow depth, inasmuch as high WTL inhibits rapid cooling and freezing of the peat column. These features of the ridge also affect the timing of the lowest temperatures onset. As compared to other sites (hollow, fen, low and high ryam) where the minima in the annual course of soil surface temperatures (–0.8, –1.7, –2.5 and –2.4 °C, respectively) are reported in February, the temperature minimum on the ridge of RHC is commonly observed in January (–5.2 °C).

The minima in monthly mean soil temperatures in the period from 2011 to 2017 were primarily observed in February in the 0–20 cm layer, however in some years they were reported in January or December. Over the entire observation period, minimum soil temperatures constituted –5.7 °C on the surface (high ryam, January 2013), and –2.3 °C at a depth of 20 cm (high ryam, February 2012).

The differences in temperatures at equivalent depths between the plots tend to take the highest values (>5 °C) at the 30–60 cm depth during the warm season. On average, the differences do not exceed 1.8 °C on a yearly time-scale. Forested areas with low WTL appear to be colder throughout the profile. The smallest differences in average annual (<0.5 °C) and maximum (<0.8 °C) annual course of soil temperatures along the soil profile are observed between sites with close WTLs. Besides, the maximum temperatures reported from the ridge of RHC occupying an intermediate position between these groups, are closer to the maxima observed at plots with high WTL. The differences in maximum temperatures among the plots tend to increase with depth, reaching 4 °C within the 30–40 cm depth interval, at which differences between the annual minima in soil temperatures are the lowest (±0.5 °C) for all the investigated plots. Due to the presence of mineral soil already at a depth of 100 cm in the high ryam plot, the minimum soil temperature at depths below 40 cm is lower there by 1.6–1.8 °C, against other plots, where peat layer thickness reaches 210–310 cm.

Previously, the authors showed that peat soil in low ryam exhibited a smoother temperature dynamics, as compared to mineral soil [*Dyukarev and Golo-*

vatskaya, 2013]. The mean monthly data indicate that during the warm season the upper 80 cm soil layer of the low ryam is colder by 5–7 °C, than the mineral soil plot, and, conversely, in the winter season it is warmer by 0.3–1.0 °C. This temperature difference between mineral and peat soils persisted (although less pronouncedly) in plots with high WTL (fen and hollow of RHC). The enhanced thereby thermal inertia of the peat deposit largely decelerates both its heating and cooling.

SEASONALLY FROZEN LAYER

October–November is the time when the inception of a seasonally frozen layer occurs in the upper peat layers across all the plots. A decrease in the air temperature coupled with the atmospheric precipitation falling in the form of snow, which is typical of these months, paves the way for the formation of permanent snowpack in the subsequent period. The differences in the dates of the surface temperature transition through 0 °C threshold towards negative values can be observed during a few days across the investigated plots. During the study period, the dates of surface temperature transition through 0 °C threshold, to become zubzero, occurred on October, 26 (fens), October, 28 (hollow of RHC), October, 31 (ridge of RHC and high ryam), November, 4 (low ryam). The earliest onset of soil freezing associated with the temperatures achieving steady negative values was reported in 2016, specifically: October, 12 (hollow and fen), October, 13 (ridge) and October, 18 (low and high ryams). In 2011 and 2013, the shift to subzero soil temperatures occurred as late as November, 7–10. The timing of the formation of permanent snowpack, its depth and bog water table level in the beginning of winter largely affect the temperature minima within the annual course of soil temperature (Fig. 3). The interplay of severe frosts in November 2015 (22.11.2015 air temperature dropped to –22.8 °C) and low snowpack height (14 cm) facilitated deeper frost penetration: up to 31 cm in the high ryam plot, which is 5–20 cm more, than in other years. Given that the snowpack height during the 2012/13 winter was the lowest for the considered period and varied from 20 cm in December to 30 cm in March, the deepest frost penetration (up to 61 cm) during the study period was observed in the high ryam plot in March 2013. In the winter of 2014/15, the snowpack whose arrival was reported on November 6, had increased up to 103 cm by March 14. Whereas bog soils were frozen only to a depth of 18–32 cm, even when affected by severe frosts (–34 °C).

The maximum depth of peat column freezing was observed in the period from February through April (Table 1). Both the date of its achieving and penetration magnitude are largely controlled, besides the snowpack availability, by a set of meteorological parameters of a specific year: the dates of early frost; air temperature; duration of the cold spell with low snow depth or snow-free period; snowpack height and specific conditions of its formation at different plots.

The maximum depth of soil freezing for most of the investigated plots was observed during the 2011/12 moderately cold winter, reaching 34–61 cm (Fig. 3). The exception is the ridge of RHC, where the maximum depth of frost penetration constituted

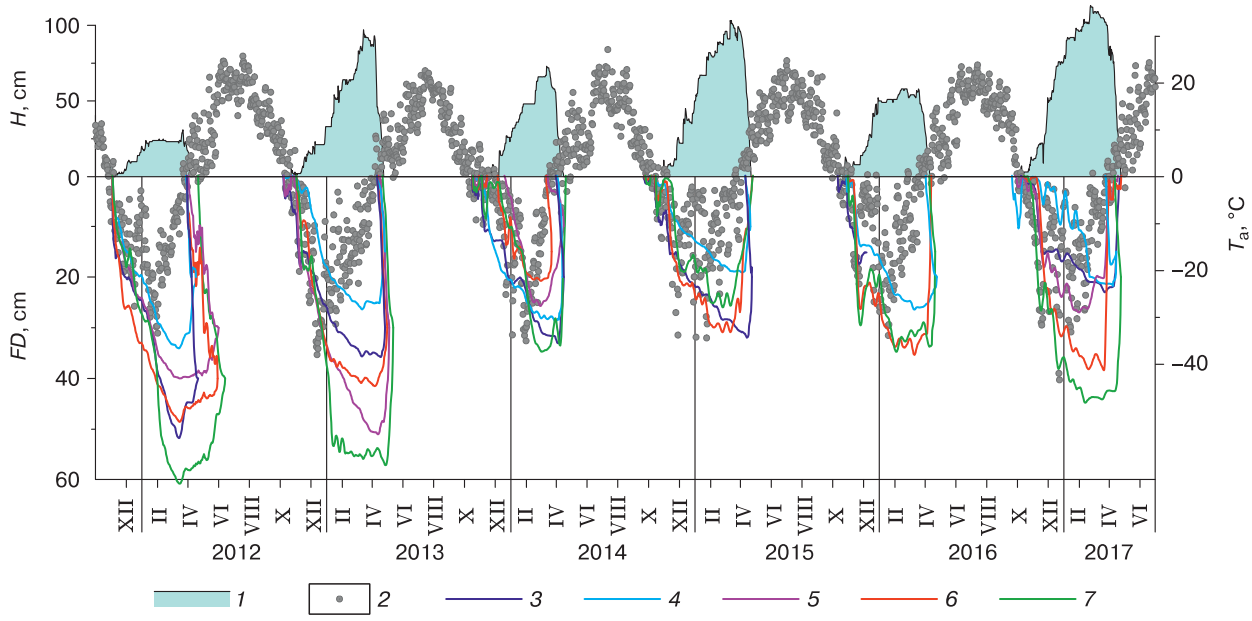

Fig. 3. Variations of freezing depth *FD*, snowpack height *H* (*1*) and air temperature T_a (*2*).

3 – open fen; *4* – hollow of RHC; *5* – ridge of RHC; *6* – low ryam; *7* – high ryam.

Table 1. **Maximum freezing depth of peat deposit (*FD*, cm) and the date of its achieving**

Plot	2011/12		2012/13		2013/14		2014/15		2015/16		2016/17		Mean ±SD	
	Date	*FD*	Date	*FD*	Date	*FD*	Date	*FD*	Date	*FD*	Date	*FD*	Date	*FD*
Open fen	15.03	**52**	18.04	36	05.04	33	14.04	32	–	–	23.03	23	02.04 ± 14	35 ± 11
Hollow of RHC	15.03	**34**	06.04	25	31.03	29	03.04	19	10.03	26	08.04	22	27.03 ± 12	26 ± 5
Ridge of RHC	16.03	40	13.04	**51**	01.03	26	–	–	–	–	01.02	27	08.03 ± 30	36 ± 12
Low ryam	15.03	**49**	06.04	42	02.03	21	24.02	31	10.03	35	21.03	38	13.03 ± 15	36 ± 10
High ryam	15.03	**61**	28.04	57	03.03	35	25.02	26	04.02	35	12.02	45	05.03 ± 30	43 ± 14

Note. Bold type indicates maximum freezing depth in each plot. Dash shows missing data (n/a), SD is standard deviation.

51 cm at the end of the 2012/13 winter. The lowest frost penetration depth was 19–26 cm during the winters of 2014/15 (hollow of RHC and high ryam), 2016/17 (ridge and open fen) and 2013/14 (low ryam).

The peat soil freezing from the surface downward to the maximum depth occurs at a rate of 0.2–0.3 cm/day. The rate of freezing achieved maximum (0.51 cm/day) in 2012 in the high ryam plot (Fig. 3). However, by dividing the frozen layer into two parts – shallow and deeper – we can see that at depths up to 20 cm from the surface, the freezing proceeds more intensely (at a rate averaging 0.60–0.70 cm/day), than in the lower parts. The maximum rate of freezing may differ significantly from the mean values. Thus, it reached 1.43 cm/day on the ridge in 2011, and 1.67 cm/day in the high ryam plot in 2015. The rate of freezing in wooded areas is generally higher, than in open, windswept areas, which is largely explained by the bog water level. At depths below 20 cm, the rate of freezing averages 0.21 cm/day. The maximum rate (0.81 cm/day) was reported in the high ryam plot in 2014. The freezing rate slows down in the underlying layers largely due to a thicker snowpack and slower cooling of the peat layer, which at this point is warmer, than the overlying substrate.

Degradation of the seasonally frozen layer begins both from above and below, which is prompted by a relatively warm, permanently unfrozen peat layer sitting below the active layer. Given that the snowpack effect is the most prominent from mid to late winter, it largely mitigates the action of negative air temperatures on low temperatures in the seasonally frozen layer, allowing thereby for the thaw process to begin in the peat column from below upwards. However, given that initially the rate of the active layer thawing from below is exceedingly low, its degradation therefore occurs most rapidly from above (Fig. 3).

The average date of the beginning of the seasonally frozen layer melting from above varies significantly from year to year, as well as between the investigated plots. The earliest initiation of the active layer thaw (15.03.2014) was reported from the low ryam plot, and the latest date (26.04.2013) from the high ryam plot. The rate of thawing from above varies from 1–5 to 30–35 cm/day. In some years, the frozen soil layer degradation began first from below, for example, in 2014 and 2015 (high tall ryam), and in 2015, 2016 and 2017 (low ryam) (Fig. 3).

Over the entire observation period, the longest duration (217 days) of the seasonally frozen layer was observed in the high ryam plot in the winter of 2011/12, and the shortest (122 days) in the low ryam plot in the winter of 2013/14. The average duration of the seasonally frozen state of soils was 176 days for all the observation plots. In the 2011/12 winter, the longest frozen-state period (201–217 days) was observed at sites with low WTL (high/low ryams and ridge of RHC). In 2014/15 and 2016/17, its duration showed the least variations (183–191 days) across the observation plots. The differences in the frozen layer duration between the plots (cumulative variations in individual years) are seen to be less, than variations within one site in different years, suggesting whereby the interplay of the same factors affecting the dates of the beginning and end of the frozen-state of soils.

EVALUATION OF THE SNOWPACK AND AIR TEMPERATURE EFFECTS ON MAXIMUM FROST PENETRATION DEPTH

During the cold season, the snowpack viewed as effective thermal insulator largely precluding soil cooling, plays a critical role in the formation of the seasonally frozen layer. Snowpack reduces the amplitude of fluctuations of the surface soil temperature and largely controls the maximum depth of negative temperatures penetration into the peat column [*Kudryavtsev, 1981, p. 82–110; Buldovich and Ershov, 2001; Gavriliev, 2004*].

The warming effect of the snowpack during the winter season is material only at its small height [*Sherstyukov, 2008*]. The insulating capacity (thermal resistance) is determined by the thermal conductivity of snowpack [*Kudryavtsev, 1954; Buldovich and Ershov, 2001; Ershov, 2004, p. 248–303; Osokin et al., 2013*]. Moss cover significantly affects the soil thermal regime during the warm season [*Garagulya and Ershov, 2001*], while in the cold season, its influence is easily arrested by the progression of cooling [*Tishkov et al., 2013; Porada et al., 2016*]. A year-round warm,

lower part of the peat column also plays an important role in the formation of the seasonally frozen layer. When freezing of water-saturated peat layers proceeds slowly, enormous heat is released during phase change of water [*Kudryavtsev, 1981, p. 24–45*]. Given that the snowpack shields this heat, the soil freezes to a shallow depth [*Kiselev et al., 2017*].

In the permafrost science, the seasonal freeze/thaw depth is determined either by different approximate formulas (e.g. offered by L.S. Leibenzon, D.V. Redozubov, V.S. Luk'yanov, V.A. Kudryavtsev, M.D. Golovko, A.V. Pavlov, etc.) [*Kudryavtsev, 1981, p. 24–45; Ershov, 2004, p. 248–303*] or by solving the system of equations of heat and moisture transfer with phase transitions in non-stationary conditions [*Chudnovsky, 1976; Buldovich, 2001; Arkhangelskaya, 2012*].

The authors have proposed a regression model for approximate prediction of the freezing depths and potential changes in the permafrost conditions within the study area

$$FD_{\max} = a_0 + a_1 T_a + a_2 H_{\max},$$

where $FD_{\max}$ is maximum freezing depth during the winter season, cm; T_a is the winter mean air temperature, °C; $H_{\max}$ is the maximum snowpack height, cm; a_0, a_1, a_2 are the model coefficients.

Among the control temperatures are also mean air temperatures in individual months (September through March) and mean temperatures for periods spanning several months. The best results were achieved for the December–February mean air temperatures. The involvement of soil water content (or bog water level) as a proxy control parameter can improve the model accuracy [*Kudryavtsev, 1981, p. 24–45; Ospennikov, 2001*], although all the study plots are water-logged and fluctuations of water content in the uppermost peat layer reach 80–100 %.

Table 2. **Model coefficients (a_0, a_1, a_2), average module of mean absolute error (*MAE*) and coefficient of determination (R^2)**

Investigated plot	a_0, cm	a_1, cm/°C	a_2, cm/cm	*MAE*, cm	R^2
Open fen	51.57	–0.45	–0.29	3.41	0.85
Hollow of RHC	25.86	–0.72	–0.15	1.07	0.93
Ridge of RHC	–70.76	–6.01	0	1.65	0.96
Low ryam	41.68	–3.69	–0.42	1.69	0.94
High ryam	–13.69	–4.20	–0.14	3.29	0.88

Analysis of the field observations results using the STATISTICA software (Statsoft/Dell, USA) enabled evaluation of the model coefficients (Table 1) for each plot, as well as the model calculation error and the coefficient of determination for a linear regression model (Table 2). The results revealed that the proposed model has best described the observed variations of the maximum freezing depth for RHC and low ryam plot (Fig. 4, Table 2).

The cumulative forecasting error, on average, does not exceed 1.7 cm; the determination coefficient varies from 0.63 to 0.96. The model results are slightly worse for open fen and high ryam: the forecasting error averaged 3.4 and 3.3 cm, and the determination coefficient was 0.85 and 0.89, respectively. The largest deviation (8.5 cm) of model values from the observed values was obtained in the high ryam plot in the winter of 2013/14, which is probably associated with the specific pattern of snow accumulation during that year. In other years, the forecasting error is <5 cm for all the plots.

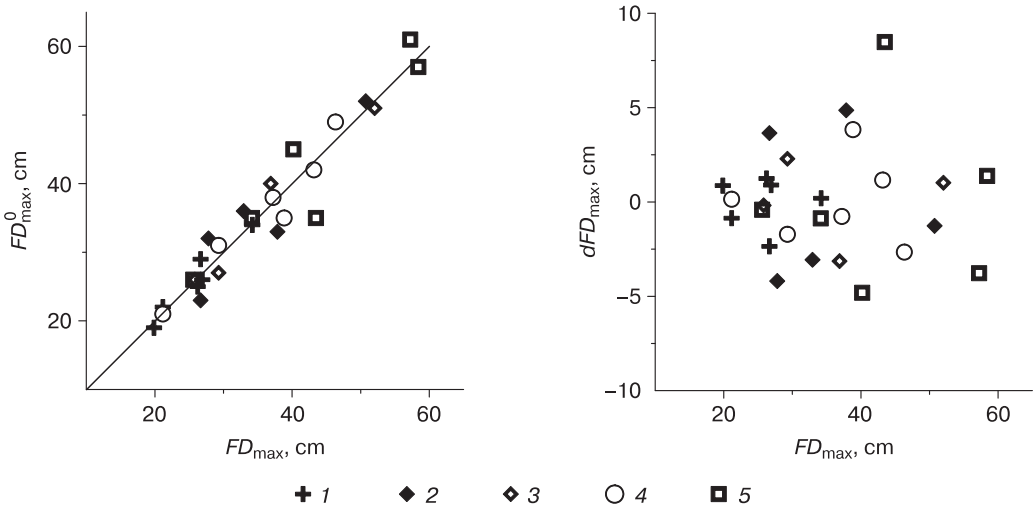

Fig. 4. Comparison of modeled maximum freezing depth values ($FD_{\max}$) with measured ($FD^0_{\max}$) and calculation arrow ($dFD_{\max} = FD_{\max} - FD^0_{\max}$):

1 – open fen; *2* – hollow of RHC; *3* – ridge of RHC; *4* – low ryam; *5* – high ryam.

CONCLUSIONS

Based on the ASMC (Atmospheric-soil measuring complex) measurements, some specific features of the temperature regime of peat soils in oligotrophic bog have been determined for the southern taiga zone of Western Siberia.

The variations of soil temperature and seasonal freeze depth are found to be significant for the plots with high and low bog water levels. Given close air temperatures, the plots with high WTL warmed faster during the warm season, as compared to plots with low WTL. At the plots with low WTL, the tree-level stand provoked a reduction in the incident solar radiation, whereas the loose topmost layers of mossy mat acted as a heat-insulating agent precluding the peat column warming. At this, high temperature gradients were observed in the moss surface layer. The oversummer soil temperature was lower in the waterlogged plots, than in non-waterlogged areas with mineral soil, due to the additional energy losses consumed by evaporation. The differences in temperature gradients at the same depths between the plots did not exceed 1.8 °C, on average, however during the warm season they reached 3.2–6.4 °C in the 10–80 cm depth interval.

The formation of the seasonally frozen layer is reported to begin on average after October 26 in the studied peatland during the period 2011–2017. The frost penetration depth was the highest during the winter season in soils of water-logged areas (26–35 cm), and in areas with low WTL (36–43 cm). The thickness of seasonally frozen layer in mineral soils averaged 93 cm (maximum value: 144 cm). The water-saturated peat layers prevented deeper freezing due to the heat released during the phase change of water which was shielded by the thick snowpack causing thereby the peat soil freeze to a shallow depth. The proposed regression model, involving the maximum snow depth and the winter mean air temperature as key characteristics, can be used for a rough estimation of the maximum freezing depth.

The duration of the seasonally frozen layer for different sites varied from 122 to 217 days, averaging 176 days. Differences in the duration of the frozen-state between sites in the specific year of study is less than the interannual differences at each site, indicating a single set of factors affecting the dates of initiation and end of freezing. The maximum freezing depth, on the contrary, showed greater variability, when individual sites were compared, i.e. this parameter was more appreciably affected by local geomorphological settings (peat density, WTL, vegetation cover), as compared to the interannual variability of the weather conditions.

The annual course of soil temperature of the considered peatland ecosystems appears smoother in comparison with plots with mineral soil. The upper layers of peat soil are cooler during the warmer months, and warmer in the cold season, as compared to mineral soil. The enhanced thermal inertia of the peat deposit prevents both its heating and cooling. The depth of freezing in the bog environment is 1.5–3-lower, than in mineral soils [*Dyukarev and Golovatskaya, 2013*]. The relic signs of cryogenesis in the margins of the studied peatland [*Dyukarev and Pologova, 2007*] indicate the cryogenic phenomena to have been wider spread in the past. The ongoing climate change manifests itself in the shift of boundaries of climatic zones and permafrost zones. Note that early in the last century the southern taiga of Western Siberia was within the zone of island permafrost distribution, while at the end of the century it is found to be located within the bounds of the zone of long-term seasonally freezing soils [*Dyukarev and Pologova, 2007*].

Obtaining new field data on the modern thermal and hydrological regimes of different types of wetland ecosystems in Western Siberia allows developing land surface models, while calibration of these models based on the observational data will improve understanding of the role of wetland ecosystems in the formation of regional climate and clarify parameterization of the terrestrial ecosystems in modern climate models.

The work was financially supported by the Russian Foundation for Basic Research (Project # 18-05-00306) and YuSU grant (# 13-01-20/39).

References

Anisimov, O.A., Borsch, S.V., Georgievskii, V.U., et al., 2012. Methods for assessing the effects of climate change on physical and biological systems. Planeta, Moscow, 512 pp. (in Russian)

Anisimov, O.A., Kokorev, V.A., 2017. Russian permafrost in the 21st century: model-based projections and analysis of uncertainties. Earth's Cryosphere XXI (2), 3–9.

Arkhangelskaya, T.A., 2012. Temperature regime of a complex soil cover. GEOS, Moscow, 281 pp. (in Russian)

Barid, A., Belyea, L., Comas, X., et al., 2013. Carbon Cycling in Northern Peatlands. Geophys. Monograph Ser., AGU, vol. 184, 297 pp.

Boch, M.S., Mazing, V.V., 1979. Mires ecosystems of the USSR. Nauka, Moscow, 188 pp. (in Russian)

Bukvareva, E.N., 2010. The role of terrestrial ecosystems in climate regulation and Russia's place in the post-Kyoto process. KMK Publ. House, Moscow, 97 pp. (in Russian)

Buldovich, S.N., 2001. Heat transfer and temperature formation of rocks. In: E.D. Ershov (Ed.). Fundamentals of Geocryology. Part 4. Dynamic Geocryology. Moscow University Press, Moscow, pp. 93–107. (in Russian)

Buldovich, S.N., Ershov, E.D., 2001. The influence of snow cover on the formation of temperature regime of rocks. Ibid., pp. 50–64. (in Russian)

Bulygina, O.N., Veselov, V.M., Razuvaev, V.N., Aleksandrova, T.M., 2017. Description of an array of current data on key meteorological parameters at stations in Russia. Certificate of state registration of database No. 2014620549, http://

meteo.ru/data/163-basic-parameters (submittal date: 12.12.2017).

Chechkin, S.A., 1970. Water-thermal regime of undrained marshes and its calculation. Gidrometeoizdat, Leningrad, 205 pp. (in Russian)

Chigir, V.G., 1978. Thermal Melioration of Long-term Seasonally Frozen Soils. Nauka, Moscow, 147 pp. (in Russian)

Chudnovsky, A.F., 1976. The Heat Physics of Soil. Nauka, Moscow, 353 pp. (in Russian)

D'Angelo, B., Gogo, S., Laggoun-Défarge, F., et al., 2016. Soil temperature synchronization improves representation of diel variability of ecosystem respiration in Sphagnum peatlands. Agricultural and Forest Meteorol. 223, 95–102.

Desyatkin, R.V., Desyatkin, A.R., 2017. Temperature regime of solonetzic meadow-chernozemic permafrost-affected soil in a long-term cycle. Eurasian Soil Sci. J. 50 (11), 1301–1310.

Desyatkin, R., Desyatkin, A., Fedorov, A., Konstantinov, P., 2015. Air temperature changes and their impact on permafrost ecosystems in Eastern Siberia. Thermal Science 19, S351–S360.

Dokturovsky, V.S., 1932. Peatlands. Course of lectures on wetland sciences. Gos. Nauchno-Tekhnicheskoe Gornoe Izd-vo, Moscow, Leningrad, 192 pp. (in Russian)

Drozdov, O.A. (Ed.), 1957. Methods of climatological processing of meteorological observations. Gidrometeoizdat, Leningrad, 492 pp. (in Russian)

Dyukarev, E.A., 2015. Influence of air temperature and snow cover on the seasonally frozen soil layer characteristics. Earth's Cryosphere XIX (3), 39–45.

Dyukarev, E.A., Golovatskaya, E.A., 2013. Specific features of the temperature regime of peat deposit of oligotrophic mire of southern taiga of Western Siberia. Geografiya i Prirodnye Resursy (Geography and Natural Resources), No. 1, 65–71.

Dyukarev, E.A., Golovatskaya, E.A., Duchkov, A.D., Kazantsev, S.A., 2009. Temperature monitoring in Bakchar bog (West Siberia). Russian Geology and Geophysics 50 (6), 745–754.

Dyukarev, A.G., Pologova, N.N., 2007. Modern cryomorphoses in landscapes of the southern taiga of Western Siberia. Geografiya i Prirodnye Resursy (Geography and Natural Resources), No. 1, 96–100.

Dyukarev, E.A., Pologova, N.N., Dyukarev, A.G., Golovatskaya, E.A., 2011. Forest cover disturbances in the south taiga of Western Siberia. Environ. Res. Lett. 6 (3), 035203 (9 p).

Efremov, S.P., Efremova, T.T., Melent'eva, N.V., 1994. Carbon reserves in bog ecosystems. In: Carbon in the Ecosystems of Russian Forests and Bogs. Computing Center SB RAS, Krasnoyarsk, pp. 128–139. (in Russian)

Ershov, E.D. (Ed.), 1989. Geocryology of the USSR. Western Siberia. Nedra, Moscow, 454 pp. (in Russian)

Ershov, E.D. (Ed.), 2004. Methods of Geocryological Studies. Moscow University Press, Moscow, 512 pp. (in Russian)

Evseeva, N.C., Petrov, A.I., Kuzhevskaya, I.V., Kharanzhevskaya, Yu.A., 2016. Characteristics of the snow cover in Tomsk region. Geography and Environmental Management of Siberia, No. 21, 56–70. (in Russian)

Garagulya, L.S., Ershov, E.D., 2001. Influence of vegetation on the temperature regime of rocks. In: E.D. Ershov (Ed.). Fundamentals of Geocryology. Part 4. Dynamic Geocryology. Moscow University Press, Moscow, pp. 64–75. (in Russian)

Gavriliev, R.I., 2004. A Guide on Thermophysical Properties of Environmental Components in the Permafrost Zone. Publ. House SB RAS, Novosibirsk, 146 pp. (in Russian)

Gilichinsky, D.A., 1986. Seasonal Cryolithozone of Western Siberia. Nauka, Moscow, 143 pp. (in Russian)

Golovatskaya, E.A., Dyukarev, E.A., 2012. The influence of environmental factors on the CO_2 emission from the surface of oligotrophic peat soils in West Siberia. Eurasian Soil Science J., No. 6, 658–667.

Golovatskaya, E.A., Dyukarev, E.A., Ippolitov, I.I., Kabanov, M.V., 2008. The impact of landscape and hydrometeorological conditions on CO_2 emission from peat bog ecosystems. Doklady RAN (Proceedings of the Russian Academy of Sciences), No. 4, 1–4.

IPCC, 2013. Climate Change 2013: The Physical Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental. Panel on Climate Change. T.F. Stocker, D. Qin, G.-K. Plattner et al. (Ed.). Cambridge University Press, Cambridge, New York, 1535 pp.

Ivanov, K.E., Novikov, S.M., 1976. Peatlands of Western Siberia, their Structure and Hydrological Regime. Gidrometeoizdat, Leningrad, 447 pp. (in Russian)

Kabanov, M.V., 2015. Regional climate regulation factors in Western Siberia. Geografiya i Prirodnye Resursy (Geography and Natural Resources), No. 3, 107–113.

Kalyuzny, I.L., Batuev, V.I., 2015. Formation of temperature regime of peat deposit under changing climatic characteristics in the northern and northwestern zone of oligotrophic bogs in the European part of Russia. In: Proceedings of the Main Geophysical Observatory named after A.I. Voeikov, No. 577, pp. 156–168. (in Russian)

Karelin, D.V., Zamolodchikov, D.G., 2008. Carbon Exchange in Cryogenic Ecosystems. Nauka, Moscow, 344 pp. (in Russian)

Kiselev, M.V., Voropay, N.N., Dyukarev, E.A., 2017. Temperature of soil in the sedge-sphagnum mire of the raised bog in the southern taiga of Western Siberia. Geografiya i Prirodnye Resursy (Geography and Natural Resources), No. 3, 110–117.

Kitaev, L.M., Kislov, A.V., 2008. Regional differences of snow accumulation: contemporary and future changes (on the example of Northern Europe and northern West Siberia). Kriosfera Zemli (Earth's Cryosphere), XII (2), 98–104.

Kudeyarov, V.N., Zavarzin, G.A., Blagodatskii, S.A., et al., 2007. Carbon pools and fluxes in terrestrial ecosystems of Russia. Nauka, Moscow, 315 pp. (in Russian)

Kudryavtsev, V.A., 1954. Temperature of the upper horizons of permafrost within the USSR. Izd-vo AN SSSR, Leningrad, 182 pp. (in Russian)

Kudryavtsev, V.A. (Ed.), 1981. Permafrost Studies (short course). Moscow University Press, Moscow, 240 pp. (in Russian)

Kurakov, S.A., 2012. Environmental condition autonomous monitoring system. Datchiki i Sistemy (Sensors and Systems), No. 4, 29–32.

Liss, O.L., Abramova, L.I., Avetov, N.A., et al., 2001. Mire Systems of Western Siberia and their Conservation Value. Grif and K°, Tula, 584 pp. (in Russian)

Osokin, N.I., Sosnovsky, A.V., 2014. Spatial and temporal variability of the thickness and density of snow cover in Russia. Led i Sneg (Ice and Snow), 128 (4), 72–80.

Osokin, N.I., Sosnovsky, A.V., 2015. Effect of air temperature and snow depth dynamics on frost depth. Earth's Cryosphere XIX (1), 88–93.

Osokin, N.I., Sosnovsky, A.V., Nakalov, P.R., Nenashev, S.V., 2013. Thermal resistance of snow cover and its effect on soil freezing. Led i Sneg (Ice and Snow), 121 (1), 93–103.

Ospennikov, E.N., 2001. Effect of marsh-ridden areas on the temperature regime of underlying rocks. In: E.D. Ershov

(Ed.). Fundamentals of Geocryology. Part 4. Dynamic Geocryology. Moscow University Press, Moscow, pp. 85–93. (in Russian)

Pavlov, A.V., 2008. Monitoring of the Cryolithozone. Academic Publishing House "Geo", Novosibirsk, 230 pp. (in Russian)

Peng, X., Zhang, T., Frauenfeld, O.W., et al., 2017. Response of seasonal soil freeze depth to climate change across China. The Cryosphere, No. 11, 1059–1073.

Porada, P., Ekici, A., Beer, C., 2016. Effects of bryophyte and lichen cover on permafrost soil temperature at large scale. The Cryosphere, No. 10, 2291–2315.

Romanov, V.V., 1961. Wetlands Hydrophysics. Gidrometeoizdat, Leningrad, 359 pp. (in Russian)

Rydin, H., Jeglum, J., 2015. The Biology of Peatlands. Oxford University Press, Oxford, 400 pp.

Semenov, S.M., 2004. Greenhouse Gases and Present Climate of the Earth. ITs Meteorologia i Gidrologia, Moscow, 175 pp. (in Russian)

Serebryanskaya, G.A., 1946. Freezing and thawing of soil in the central part of Baraba. Pochvovedeniye (Eurasian Soil Science), No. 9, 555–564.

Sheng, Y., Smith, L.C., MacDonald, G.M., et al., 2004. A high-resolution GIS-based inventory of the west Siberian peat carbon pool. Global Biogeochem. Cycles 18, GB3004.

Sherstyukov, A.B., 2008. Correlation of soil temperature with air temperature and snow cover depth in Russia. Kriosfera Zemli (Earth's Cryosphere), XII (1), 79–87.

Sherstyukov, A.B., Sherstyukov, B.G., 2015. Spatial features and new trends in thermal conditions of soil and depth of its seasonal thawing in the permafrost zone. Russian Meteorology and Hydrology 40 (2), 73–78.

Stolbovoi, V., McCallum, I., 2002. CD-ROM Land Resources of Russia. Laxenburg, Austria, Interna. Inst. for Applied Systems Analysis and the Russian Academy of Science. CD-ROM 2002. – URL: http://www.iiasa.ac.at/Research/FOR/russia_cd/guide.htm (submittal date: 12.02.2019).

The Second Roshydromet Assessment Report on climate change and its consequences on the territory of the Russian Federation, 2014. Moscow, Federal Service for Hydrometeorology and Environment Monitoring, 61 pp. (in Russian)

Tishkov, A.A., Osokin, N.I., Sosnovsky, A.V., 2013. The effect of moss sinusia on the arctic active layer. Izv. RAN. Ser. geogr. (Bulletin of the Russian Academy of Sciences. Geographical series), No. 3, 39–46.

Trofimova, I.E., Balybina, A.S., 2015. Regionalization of the West Siberian Plain from the thermal regime of soils. Geografia i Prorodnye Resursy (Geography and Natural Resources), No. 3, 234–244.

Vasil'chuk, Yu.K., 2013. Modern southern limit of permafrost in Western Siberia Lowland. Kriosfera Zemli (Earth's Cryosphere), XVII (1), 17–27.

Vomperskii, S.E., 1968. Biological basis of the effectiveness of forest drainage. Growth of forest stands in connection with the most important environmental factors of drained peat soils. Nauka, Moscow, 312 pp. (in Russian)

Vomperskii, S.E., Sirin, A.A., Tsiganova, O.P., et al., 2005. Marshes and wetlands of Russia: an attempt to analyze the spatial distribution and diversity. Izv. RAN. Ser. geogr., No. 5, 39–50.

Voropay, N.N., Vlasov, V.K., 2017. Characteristics of the snow cover distribution on the coast of Lake Baikal. Led i Sneg (Ice and Snow), 57 (3), 355–364.

Wang, K., Zhang, T., Zhong, X., 2015. Changes in the timing and duration of the near-surface soil freeze/thaw status from 1956 to 2006 across China. The Cryosphere, No. 9, 1321–1331.

Wegmann, M., Orsolini, Y., Dutra, E., et al., 2017. Eurasian snow depth in long-term climate reanalyses. The Cryosphere, No. 11, 923–935.

Zhong, X., Zhang, T., Kang, S., et al., 2018. Spatiotemporal variability of snow depth across the Eurasian continent from 1966 to 2012. The Cryosphere, No. 12, 227–245.

Received March 11, 2018
Revised version received March 11, 2019
Accepted April 15, 2019