Temperature regimes of drained and natural peatlands in arid and water-logged years

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Abstract. The results of a study of temperature regimes of oligotrophic bog ecosystems in the south taiga zone of Western Siberia in 2011-2018 are presented. Soil temperature regimes are studied using an atmospheric-soil measuring complex at several depths from the surface to 320 cm. Waterlogged (excessive moisture) and arid years are determined by calculating the aridity index of Ped (S). The years with most severe hydrothermal conditions over the past 10 years have been found to be 2012 and 2018. In 2012, a drought is observed from April to July, with extreme aridity in June (S = 4.07), and a severe drought in July (S = 3.42). It has been found that the native treed bog in a wet year warms up much more than in a dry year. On average, the temperature values are higher by 1.5°C for all depths. On average, the drained bog is colder than the natural one at depths up to 80 cm by 4.0°C, and at depths of 120 to 240 cm by 2.5°C. The maximum temperature differences of the native bog during the warm period of a waterlogged (excessive moisture) year are 8.0°C (at a depth of 60 cm) and, at depths from 80 to 240 cm they are reduced from 5.0 to 2.5°C. The maximum differences of the native bog in the waterlogged (excessive moisture) 2018 and dry 2012 in the warm period at a depth of 60 cm are 8.0-9.0°C, and from 80 to 240 cm they are reduced from 6.0 to 1.5°C.

1. Introduction

A peat deposit is a complex organo-mineral system with specific properties: high water content and porosity, with a large amount of slightly decomposed organic matter [1]. The temperature of the peat soil affects the growth of terrestrial vegetation (mosses, herbs, shrubs, etc.), the formation of a microclimate, and is a key factor controlling many biotic and abiotic processes in soils (peat or mineral): the decomposition and mineralization of soil organic matter, greenhouse gas emissions (CO₂, CH₄, N₂O), and the release of dissolved organic carbon [2].

The main world reserves of peat (and carbon) are located in Western Siberia. According to various estimates, the peatlands of Western Siberia contain up to 70 billion tons of carbon [3]. Such a significant reservoir of carbon due to climate change or anthropogenic impact can potentially transfer to the atmosphere in the form of CO₂ or CH₄ and make a significant contribution to the carbon balance of the atmosphere.

Studies of the temperature regime of soils were carried out in the territory of the Vasyuganie geophysical station [4]. Observations were made at two sites with significantly different hydrological regimes and the levels of bog waters.
2. Study area
The first site is located 200 m from the edge of the bog massif in the basin of the Key creek. The vegetation is represented by an oligotrophic low ryam, a pine-shrub-sphagnum community with a depressed tree stand (Pinus sylvestris f. Willkommii) of 2-m height (h); the stem diameter (D) is 3-7 cm, and the closure of crowns (P) is 0.4. Chamaedaphne calyculata dominates in the dense grass-shrub layer, and Ledum palustre, Oxyccocus microcarpus, and Eriophorum vaginatum are less abundant. Large moss hummocks (h – 30-70 cm, d – 100-150 cm) are formed by Sphagnum fuscum. S. angustifolium prevails between the hummocks. The average bog water table level (WTL) is about 25 cm below the surface. The peat deposit has a depth of 2 m. The transitional and lowland woody and woody-peat deposits are under the layer of high-grade Fuscum and Magellanicum peats.

The other site is located on the drained, burned out site at Iksinskoe bog, 50 m from the main drainage channel. At the place of the former tall ryam, single tall (h – up to 6.5 m, D – 10-20 cm) pines were preserved, abundant (P – 0.7) undergrowth from pine and birch was formed (h – 2-2.3 m; D – 1.5-5 cm), a dense (P – up to 85%) shrub layer from Ledum palustre mixed with Chamaedaphne calyculata, Vaccinium myrtillus, Andromeda polifolia, Carex globularis. Small moss hummocks formed by S. magellanicum mixed with S. angustifolium and Polytrichum strictum occupy 60% of the surface. The average WTL is 50 cm. The peat deposit has a depth of 1.2 m, and is a pine-cotton grass peat. Above the middle-laid upland pine and cotton grass peats lies 20 cm of transitional woody and Magellanicum peats.

At the sites, measurements of the soil temperature were made at 14 depths (0, 2, 5, 10, 15, 20, 30, 40, 60, 80, 120, 160, 240, 320 cm) and the air temperature at a height of 2 m using an atmospheric soil measuring complex [5] in the period from September 11, 2011 to September 10, 2018 (2384 days) with a time step of 1 hour [6, 7]. The work also used data on the air temperature and rainfall sums from observations of the nearest Bakchar weather station (30 km to the west of the experimental sites) obtained from the RIHMI-WDC archive [8]. Data on the precipitation amount in the bog were obtained using a HOBO Rain Gauge 3M precipitation gauge (Onset Computer Corp., USA) installed at the first site.

The differences between the average monthly values of the air temperature measured at the meteorological station and in the bog have a seasonal variation: from March to August they are positive (the temperature in the bog is higher, on average, by 0.2-0.7°C), and in the other months they are negative (the bog temperature is lower, on average, by 0.1-0.9°C). The differences in the precipitation amount in the winter months are negative: on average, the bog falls by 10-20 mm / month less than at the Bakchar meteorological station. During the warm months (May-September), both positive deviations (up to 78 mm, July 2017) and negative ones (up to -68 mm, August, 2011) are observed, with the average difference of -3 ... - 7 mm / month.

3. Hydrothermal conditions of the territory
The specific features of the temperature regime of soils in certain years are determined not only by the characteristics of the soils, but also by the hydrothermal conditions of a particular year. The complex characteristic of temperature and atmospheric humidification are hydrothermal coefficients, which are based on the relationship between the air temperature and precipitation amount [9, 10]. It is most correct to use normalized indicators when comparing different landscapes for revealing droughts and periods of waterlogging [11, 12]. In this paper, the determination of waterlogged and dry years was carried out by calculating the aridity index of Ped [13],

$$S_i = \frac{\Delta t_i}{\sigma_t} - \frac{\Delta r_i}{\sigma_r}$$

(1)

where $\Delta t_i = t_i - t_{norm}$ is the air temperature anomaly for the $i$ time interval, $\sigma_t$ is its standard deviation, $\Delta r_i = r_i - r_{norm}$ is the anomaly of the precipitation amount over the $i$ time period, and $\sigma_r$ is its standard deviation. The S index takes positive values during droughts and negative values during periods of overwetting. According to the classification, there are weak (1 ≤ | S | <2), moderate (2 ≤ | S | <3), strong (3 ≤ | S | <4), and extreme events (| S | ≥ 4).
Despite significant differences in the monthly precipitation data from observations at the Bakchar meteorological station and the bog observation site, due to the localization of convective precipitation (especially during the summer period), the aridity / overwetting characteristics of individual months using the Ped index are identical for the two meteorological observation sites.

In relation to the long-term period (1936-2018) during the years of microclimatic observations at the meteorological station during the growing season, both dry and overwetting periods of varying intensity were observed, ranging from 1 to 4 months. The most contrasting hydrothermal conditions over the past 10 years were 2012 and 2018. In 2012, droughts were observed from April to July, with extreme aridity in June (S = 4.07) and a drought of strong intensity in July (S = 3.42). The beginning of the warm season of 2018 was characterized by a prolonged over-wetting conditions of the territory with a maximum in May (S = -4.05) (table 1).

Table 1. Aridity index of Ped.

| Year | April | May  | June | July | August | September | October |
|------|-------|------|------|------|--------|-----------|--------|
| 2011 | 2.60  | 0.64 | 1.75 | **-3.20** | **-3.49** | 1.83      | 1.34   |
| 2012 | 1.80  | 0.44 | **4.07** | **3.42** | -1.10   | 1.56      | 0.85   |
| 2013 | 0.39  | -4.31| -1.30 | 1.33 | 0.23    | -0.89     | 0.77   |
| 2014 | 0.57  | -1.26| 0.54 | 1.30 | 1.71    | -2.25     | -2.53  |
| 2015 | -1.66 | 0.16 | 1.11 | 0.86 | 0.16    | -2.06     | -2.04  |
| 2016 | 0.38  | 0.55 | 1.54 | -1.54 | 1.85    | **3.58** | -1.64  |
| 2017 | 0.32  | -0.33| -2.14| -0.29 | 1.33    | -0.41     | 0.54   |
| 2018 | **-2.51** | **-4.05** | **-1.74** | **-1.63** | 0.56    | 1.64      | **1.01** |

4. Temperature regime of naturally and anthropogenically disturbed bogs

It was found that the natural low ryam in the wet year (2018) warms up much better than in the dry year (2012). On average, for all depths the temperature in 2018 at a low ryam is higher by 1.5°C. The drained low ryam has small maximum temperatures and generally warms worse than an undisturbed low ryam under the same conditions of atmospheric drought. On average, the drained bog is colder than the natural ones at depths of up to 80 cm by 4.0°C, and at depths of 120-240 cm by 2.5°C.

At the studied sites in different years, the number of days with temperatures above +15°C was calculated. These temperatures were observed in June-August, and the duration of the period varies depending on the depth.

The maximum number of days with a peat temperature above 15°C was observed at a low ryam in the overwetted year 2018. The duration was 57 days on the surface, 42 days at a depth of 10 cm, and 19 days at 20 cm. In the dry 2012, despite the fact that the air temperature was higher than in 2018, the number of days with temperatures above +15°C at the low ryam was less at all depths except for the surface. The duration was 57 days on the surface and 29 days per 10 cm. At a depth of 20 cm, the peat warmed up to +14.8°C. Such differences show that atmospheric aridity can reduce the duration of a period with temperatures above +15°C for 10–20 days during a warm period. This decrease is mainly due to changes in the level of bog waters. Its increase leads to an increase in the amount of heat absorbed during the daytime heating of the peat deposit and to a decrease in the rate of cooling of the peat deposit during the night period.

The number of days with a peat deposit temperature of more than +15°C at the drained low ryam in the overwetted year of 2018 is much lower than in the undisturbed bog. The duration was 16 days on the surface, 4 days at 10 cm, and below it was not observed. Such a short duration of the warm
period is associated with the initially lower level of the bog waters (60 cm on average) and with a higher rate of reduction of the WTL after intense rains in comparison with the low ryam. Such differences lead to the fact that an increase in the WTL at the drained low ryam in an overmoistened year occurs for a shorter time period, which averages 5-7 days.

To assess the degree of influence of the hydrothermal conditions on the heating of the peat deposit, we calculated the differences between the soil temperature at selected sites at depths from the surface to 320 cm (Figure 1). The results obtained show that the peat deposit under the natural pine-shrub sphagnum bog in the overmoistened year 2018 warms up better than the similar peat deposit under the drained pine shrub sphagnum bog. The maximum differences for the warm period of the overwetted year (from April to the end of September) to a depth of 60 cm are 8.0°C, and at depths from 80 to 240 cm they decrease from 5.0 to 2.5°C.

![Figure 1](image.png)

**Figure 1.** Temperature differences for peat soils. A – natural low ryam \((T_{2018} - T_{2012})\), B – natural and drained ryam \((T_{\text{low ryam } 2018} - T_{\text{drained low ryam } 2018})\).

A similar result was observed when comparing the temperature of the peat deposit at the undisturbed site in the overmoistened 2018 and in the dry 2012. The maximum differences for the warm period (from April to the end of September) to a depth of 60 cm are 8.0–9.0°C, and from 80 to 240 cm they decrease from 6.0 to 1.5°C.

5. One-dimensional heat flux model

In order to investigate the effect of periods of droughts and overmoistening on the intensity of the heat exchange processes inside the soil [14, 15], we used a one-dimensional model of heat transfer (2). The heat diffusion model describes the nonstationary temperature distribution in a homogeneous solid medium within which there is no bulk motion (advection or convection). We use the one-dimensional approach due to neglectable small heat fluxes along the horizontal x and y coordinates. The term \(-\partial G/\partial z\) is related to the net conduction heat flux into the control volume for the z-coordinate direction. Within the medium there may also be an energy source term associated with the rate of thermal energy generation. This term is represented as \(dF\).
where $G$ is the heat flux due to conductive (or molecular) heat transfer, $dF$ is the change of heat not associated with molecular thermal conductivity, but with phase transitions of water, ice, and water vapor, an explicit heat transfer with water migration, heat release during wetting, heat sources from decomposition of plant remains, etc., \( \partial T / \partial t \) is the temperature gradient over time, and $z$ is the depth, cm.

Actually, the model is very simplified and does not take into account vertical heat transfer with percolated water, turbulent and radiative heat transfer within soil pores, capillary water movement, and subsequent heat effects, and also possible horizontal heat transfer related to water movement. However, the model includes two basic soil heat fluxes: conductive heat transfer and water phase transition. The model (2) is applicable for preliminary estimation of the relation between molecular (diffusive, conductive) heat flux and heat storage changes due to other reasons.

For the layer of soil located between the depths of $z_0$ and $z_1$, the following was calculated:

1. The total change in the heat content ($\Delta S$) during $\Delta t$:

   \[
   \Delta S = \int_{z_0}^{z_1} C \frac{\partial T}{\partial t} \, dz 
   \]  
   \hspace{1cm} (3)

2. The heat flux due to molecular heat transfer:

   \[
   G = -k \frac{\partial T}{\partial z} 
   \]  
   \hspace{1cm} (4)

Having determined these two values from the observation data, it is possible to calculate the heat flux due to other reasons:

\[
F = \Delta S - G
\]  
\hspace{1cm} (5)

The temperature and temperature gradients were determined by the available observation data using finite-difference schemes:

\[
\frac{\partial T}{\partial z} \approx \frac{T_{k,i}^j - T_{k-1,i}^j}{z_k - z_{k-1}}, \quad \frac{\partial T}{\partial t} \approx \frac{T_{k,i}^{j+1} - T_{k,i}^j}{\Delta t}.
\]  
\hspace{1cm} (6)

The coefficients of heat capacity ($C$) and thermal conductivity ($k$) depend on the characteristics of the soil, where the soil is represented as a three-phase medium consisting (Figure 2) of a solid phase, the gaps between which are filled with water and air [16].

![Figure 2. Composition of an Unsaturated Soil Sample [16].](image)

The heat capacity coefficient is calculated by the equation
\[
C = c_A \rho_A A + c_W \rho_W W + c_I \rho_I I + c_{S1} \rho_{S1} S_1 + c_{S2} \rho_{S2} S_2 (1 - \Pi) S_1 + c_{S2} (1 - \Pi) (1 - S_1).
\] (7)

where \(c_A, c_W, c_I, c_{S1}, c_{S2}\) are the heat capacities of the air, water, ice, peat and clay, respectively, \(\rho_A, \rho_W, \rho_I, \rho_{S1}, \rho_{S2}\) are the densities of the air, water, ice, peat and clay, respectively, \(\Pi\) is porosity.

The thermal conductivity coefficient is calculated by the following equation:

\[
k = k_A A + k_W W + k_I I + k_S (1 - \Pi),
\] (8)

where \(k_A, k_W, k_I, k_S\) are the thermal conductivities of the air, water, ice, and soil, respectively, \(A\) is the number of pores filled with air, \(W\) is the number of pores filled with water, and \(I\) is the number of pores filled with ice in the peat deposit.

At negative temperatures all water is replaced by ice \((I = W)\). Below the level of bog waters, air \((A)\) is absent, and all pores are occupied by water \((W)\) or ice \((I)\). The thermophysical characteristics of the substances are tabulated constants (Table 2) [http://thermalinfo.ru/].

| Characteristic | Heat capacity, J/kg/K | Density, kg/m³ | Thermal conductivity, W m⁻¹ K⁻¹ |
|---------------|-----------------------|----------------|-------------------------------|
| Air           | 1005                  | 1.247          | 0.025                         |
| Water         | 4182                  | 1000           | 0.59                          |
| Ice           | 2150                  | 918            | 2.25                          |
| Peat          | 1880                  | 1000           | 0.05                          |
| Clay          | 750                   | 1700           | 1.6                           |

6. Results of one-dimensional modeling of heat flux

The dynamics of daily heat fluxes at some depths over six years are presented in Figure 3. The conductive \((G)\) heat flux in summer takes positive values, which corresponds to the transfer of heat from the warm surface layers into the soil (Figure 3). In winter, the sign of molecular heat transfer changes, the flux values are usually lower than in summer.

![Figure 3. Mean daily soil conductive heat fluxes in low ryam.](image-url)

The change in the heat content \((dS)\) of the soil in summer at a depth of more than 20 cm is positive, which means the accumulation of heat in the soil. In the surface layer, the change in the heat content
has large values, while their sign changes from positive to negative during summer. Such fluctuations mean rapid heat accumulation and sink.

The dynamics of the non-conductive heat flux \( (F) \), defined as the difference between the change in the heat content and the conductive heat flux, has a more complex structure. A large negative flux \( F \) was observed in the upper layers up to 40 cm in summer (heat consumption for evaporation) and positive fluxes, in winter (heat generation when water freezes).

The maximum conductive heat transfer \( (G) \) averaged over certain months in the low ryam (Figure 4) is observed in summer not at the surface, but in a layer of 10-30 cm (at the boundary of the bog water level). In water saturated peat layers, the temperature gradients are less and, as a result, the molecular heat transfer is not so intense. However, even on the lower layer (240 cm) the heat flux is positive from June to November.

![Figure 4](image)

**Figure 4.** Soil heat fluxes in low ryam peat deposit \((G – \text{conductive heat transfer}, dS – \text{heat storage change}, F – \text{non-conductive heat flux})\).

In the cold season (from October to March), the values of conductive heat transfer \( (G) \) are negative. The largest negative, i.e., upward heat flux was obtained in December-February in a layer of 15-30 cm (inside the frozen layer). The thermal conductivity of ice is higher than that of water, and the frozen layer conducts heat better.

The change in the heat storages \( (dS) \) at monthly averaging is maximum inside the saturated water peat deposit (30-160 cm). In the upper layers with pores filled with air and in the lower layer with dense clay and less water, the heat storages vary slightly.

The distribution profiles of the non-conductive heat flux \( (F) \) are similar to the conductive flux profiles, but in the reflected (left to right) projection. For example, the largest heat fluxes entering the soil by molecular diffusion in June in a layer of 20-40 cm practically do not change in the heat storages, but they are completely spent on the evaporation of water. In winter, relatively large fluxes of molecular heat from the soil to the surface also do not cause a significant reduction in the heat storage and are generated from heat release during the formation of ice.

Inside the water-saturated peat layers (80-160 cm) there are significant fluxes of non-conductive heat transfer \( (F) \). The molecular heat transfer here is small, and the change in the heat content is significant. Interpretation of the calculation results of the fluxes is impossible within the framework of the one-dimensional heat transfer model. Possible causes of the appearance of these fluxes are the heat
release during the decomposition of organic matter or the influx / removal of heat associated with the migration of internal bog waters.

In the peat deposits with a disturbed hydrological regime due to drainage, all above-described patterns of heat flux distribution are preserved (Figure 5).

![Figure 5. Soil heat fluxes in drained low ryam (G – conductive heat transfer, dS – heat storages change, F – non-conductive heat flux).](image)

However, due to lower water levels the maximal molecular (G) transfer layer shifts deeper and to later periods (June, depth: 80 cm). The change in the heat storages (dS) is also maximal within the peat layer, which is 110 cm here.

Peat freezing in the drained area begins in November. Since the moisture in the peat is low, within only a few days the frozen layer reaches a considerable depth, which subsequently varies only slightly during winter and spring. For example, from November 8 to November 24 in 2015, the soil froze to 33 cm, and a maximum freezing depth of 45 cm was observed on May 5, 2016. Intense freezing is reflected in the non-conductive (F) heat transfer profile in November.

In general, in the natural ryam the intensity of heat exchange is higher compared to the drained ryam. This can be traced both in fluxes of molecular heat transfer and latent heat, and is explained by a high moisture content in the soil profile of the natural ryam both in the aerated part and in the water-saturated layers. In turn, in the drained ryam the moisture content is lower as the peat density is higher, the porosity is lower and, accordingly, the peat deposit water content is lower.

7. Conclusions
The temperature and moisture of peat deposits play an important role in emission of methane and carbon dioxide. The latest results show that temperatures from +15°C to +20°C initiate the most active period of methane emission [17, 18], provided that the peat deposit is filled with water. Thus, waterlogging of the peat deposit contributes to both increasing temperatures and increasing methane emissions.

Also, an increase in moisture and temperature leads to an increase in carbon dioxide emission from the surface of treed sphagnum bogs [19]. However, the peat deposit must remain unsaturated with water. Due to an initially small WTL and relatively fast filtration of water into a drained low ryam peat deposit, a significant warming up of the peat occurs in the overmoistened year. Thus, it is possible
to increase the emissions of carbon dioxide the main part of which is emitted in the upper layers of the peat deposit [20].

The results obtained above give an assessment of the influence of atmospheric aridity on the temperature regime of forested wetlands. They show some possible negative consequences of anthropogenic or natural drainage of peat deposits, such as an increase in carbon dioxide emissions from taiga bogs in wet years. One of the possible positive consequences is a slight decrease in methane and carbon dioxide emissions in dry years. However, drainage of peat deposits leads to an increase in fire danger in dry years. The above-obtained results of one-dimensional modeling of the heat fluxes have shown that this approach has insufficient efficiency in estimating the latent heat flux.

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