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THERMAL RESISTANCE OF SNOW AS A CONTROL OF PERMAFROST STABILITY

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Spatial and temporal variations in thermal resistance of snow in the territory of Russia for the period from 2001 to 2010 has been studied and mapped from data of snow surveys. Long-term means of snow thermal resistance and freezing index for 2001–2010 were compared with the respective values for 1966–2000. The thermal resistance of snow was found out to control the thermal stability of permafrost and to vary geographically. Areas with favorable and unfavorable conditions for permafrost preservation were identified.

Climate change, permafrost, snow cover, air temperature, thermal resistance

INTRODUCTION

Snow is an important element in the atmosphere–surface–subsurface interactions, which affects freezing processes and the thermal state of shallow permafrost [*Osokin et al., 2000; Sherstyukov, 2008; Park et al., 2015*]. Variations in the thermal parameters of snow can reduce or cancel climatic warming effects in some permafrost areas [*Izrael et al., 2002; Pavlov, 2008*]. The ground temperature patterns depend on the depth of snow and on the dynamics of its accumulation [*Pavlov, 2008*]: inter-annual variations in freezing index (a total of negative air temperatures) and snow accumulation can cause more than 50 % difference in frost depth [*Osokin and Sosnovskiy, 2015*].

When discussing snow as a control of the thermal regime and stability of permafrost, it is pertinent to define the meaning of "stability". Different concepts of stability applied to geosystems, including cryogenic ones [*Solomatin, 1992*], mean it as ability of systems to keep their parameters invariable with time, to resist external loads, to recover after natural and man-made disturbance, etc. The thermal stability of permafrost is evaluated proceeding from variations in air and ground temperatures and heat flux implying equilibrium between the incoming heat the ground receives in summer and the heat it looses in winter. On the other hand, the thermal regime of permafrost and its possible responses to temperature changes affect its mechanic stability, which is crucial for engineering applications, such as piling design.

The high, medium, or low thermal stability of permafrost and its sensitivity to climate change can be estimated using a dimensionless coefficient: a ratio of ground-to-air mean annual temperatures [*Pavlov and Malkova, 2009; Malkova et al., 2011*]. Snow, among other factors, can control the spatial patterns of permafrost stability. For instance, variable snow depths influence mean annual ground temperatures in the way that permafrost in southern Yakutia remains thermally stable despite considerable climate warming [*Skachkov, 2008*].

The thermal regime of shallow permafrost depends on climate, as well as on the properties of soil and vegetation. However, the soil properties are locally stable, which makes positive and negative air temperatures and snow parameters the principal factors of permafrost thermal stability at specific points. Therefore, the trends of the permafrost thermal state depend on the interplay of heat insulation by snow and the dynamics of air temperatures, especially negative ones [*Pavlov, 2008*].

Heat exchange between cold air and permafrost depends on snow insulation: the conditions for permafrost conservation and thermal stability are favorable when the snow insulation capacity decreases faster than negative air temperatures become warmer and vice versa.

Thermal insulation by snow affects heat flux in the system "near-surface air–permafrost". Heat flux q across the snow cover is found as [*Pavlov, 1979*]

$$q = \lambda_s \, \partial T / \partial x,$$

where T is the snow temperature; λ_s is the effective thermal conductivity of snow; x is the coordinate along the snow depth.

At a quasi-stationary temperature distribution over the snow depth and approximately equal temperatures of snow surface and air, the heat flux is

$$q \approx (T_a - T_g)/R_s,$$

where T_a and T_g are the air and ground temperatures, respectively, °C; $R_s = h_s/\lambda_s$; h_s is the snow depth.

At an air temperature of –10 °C, the absolute ground surface temperature ($|T_g|$, °C) under 30–80 cm thick snow is five to ten times as low as the air temperature ($|T_a|$, °C) [*Shmakin et al., 2013*]. Then the ground surface temperature is

$$T_g = kT_a,$$

where k = 0.1–0.2,
and the heat flux is

$$q \approx (1 - k)T_a/R_s.$$

According to the latter equation, heat flux is proportional to air temperature and inversely proportional to R_s. The parameter R_s is the thermal resistance found as a ratio of snow depth (h_s) to snow thermal conductivity (λ_s), which is the main control of heat insulation capacity [*Balobaev, 1991; Solomatin, 1992*]. The effect of thermal resistance on frost depth is commensurate with that of mean temperature in the cold season [*Osokin et al., 2013a*], while the variations in mean winter temperature are times lower than possible seasonal and landscape variations in the thermal resistance R_s.

In order to assess the sensitivity of the permafrost thermal stability to climate change, we compare the spatial and temporal R_s variations with the dynamics of freezing index.

THERMAL RESISTANCE OF SNOW AND ITS EFFECT ON GROUND TEMPERATURES

Thermal resisatnce of snow (R_s) depends on its effective thermal conductivity (λ_s), but the choice of the respective empirical relationships for different types of snow is problematic. The thermal conductivity λ_s inferred commonly from average snow density (ρ_s) can vary depending on snow structure and texture [*Osokin et al., 2013b, 2014*]. The coefficient λ_s [W/(m·K)] can be found using the simplified equation by *Pavlov* [*2008*]

$$\lambda_s = \kappa\rho_s, \quad (1)$$

where ρ_s is the snow density, kg/m^3; $\kappa = 10^{-3}$ is the dimensionality. This equation is valid within the snow temperature range from –10 to –20 °C, but additional 0.04 W/(m·K) should be added to or subtracted from λ_s in the case of warmer or colder snow, respectively.

According to equation (1), snow of the depth h_s = 0.50 m and density ρ_s = 200 kg/m^3 has the thermal resistance R_s = 2.5 m^2·K/W; twice thicker and denser snow (h_s = 1.0 m and 400 kg/m^3) will have approximately the same heat insulation capacity.

The effect of snow thermal resistance (R_s) on the thermal stability of permafrost inferred from heat flux data [*Solomatin, 1992*] is as follows. The thermal resistance increase from R_s = 0.86 to 1.72 m^2·K/W reduces the heat spent on thawing (warming the base of seasonal thaw to 0 °C) by factors of 1.3–1.7 and 1.4–1.6 in the northern parts of the Yamal and Gydan peninsulas, respectively, and by 1.1–1.4 times in the forested tundra of West Siberia. On the other hand, the R_s increase from 1.72 to 3.44 m^2·K/W increases the thaw depth in the north of Yamal and Gydan from 0.52 to 1.20 and from 0.95 to 1.65 m, respectively.

The effect of snow thermal resistance on ground temperatures at the depth 320 cm in the conditions of climate change for the period from 1966 to 2010 is discussed below for the case of the Verkhoyansk weather station (WMO code 24266; 67°34′ N, 133°24′ E). The total of daily positive air temperatures (T_p), or thawing index, grows 5 °C yearly at this station, with the trend $\Sigma T_p = 5.2954y - 9092.1$ (cor-

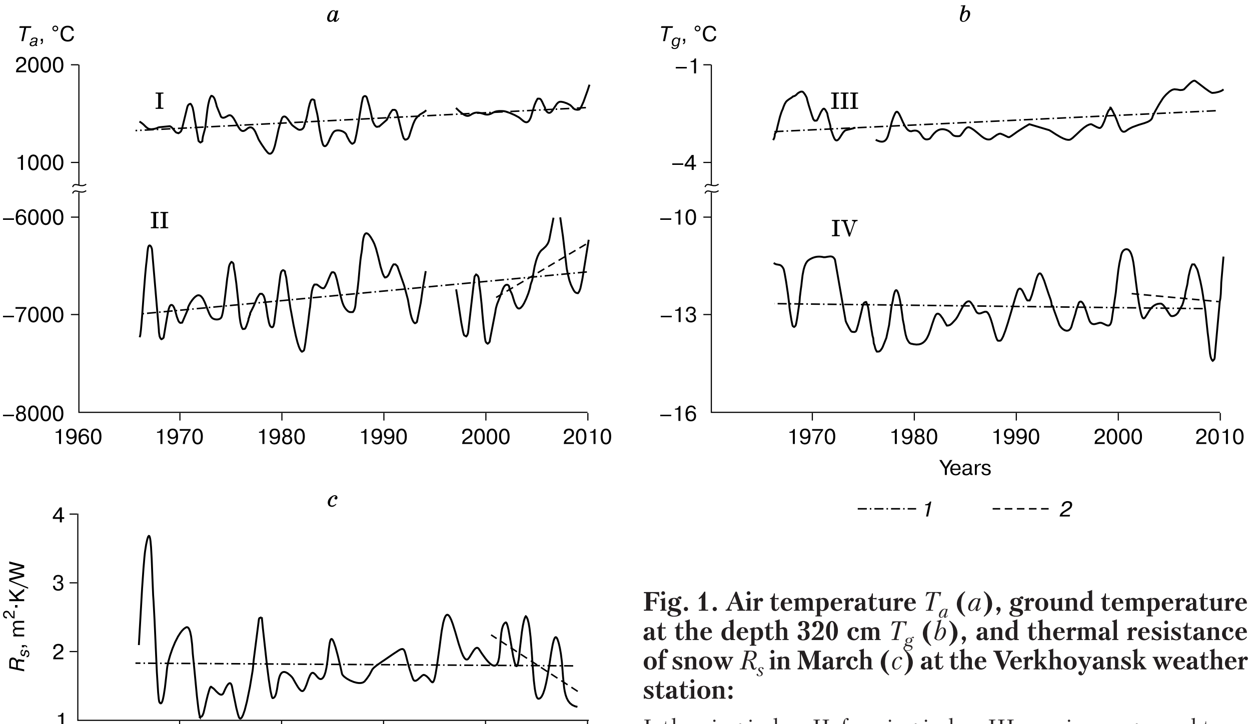

Fig. 1. Air temperature T_a (*a*), ground temperature at the depth 320 cm T_g (*b*), and thermal resistance of snow R_s in March (*c*) at the Verkhoyansk weather station:

I: thawing index; II: freezing index; III: maximum ground temperature; IV: minimum ground temperature; *1* is trend for 1966–2010; *2* is trend for 2001–2010.

relation coefficient $R^2 = 0.2078$, where y is the year from the 1966–2010 time span), or 0.4 % per year on average (Fig. 1, *a*); the total of negative daily air temperatures (T_n), or freezing index, shows an annual increase of 9 °C ($\Sigma T_n = 9.4449y - 25\,541$; $R^2 = 0.1369$), or 0.1 % on average. However, the minimum ground temperature at the depth 320 cm (Fig. 1, *b*) remains almost invariable, becoming only 0.003 °C (0.03 %) lower per year (trend: $T_g = -0.0033y - 5.9$; $R^2 = 0.0021$), due to small changes in the freezing index and thermal resistance of snow (Fig. 1, *c*). Unlike the period between 1966 and 2010, when R_s changed 0.05 % per year (trend: $R_s = 0.0009y + 0.0446$; $R^2 = 0.0005$), the freezing index from 2001 to 2010 has increased much faster to reach 59 °C or 0.9 % per year (trend: $\Sigma T_{an} = 58.658y - 124\,190$; $R^2 = 0.2831$). This, however, has not led to warming of the minimum ground temperature at 320 cm below the surface; it even decreases at a rate of 0.03 °C or 0.26 % per year (the trend: $T_g = -0.0309y + 49.678$; $R^2 = 0.0078$), as a result of considerable reduction in the thermal resistance of snow: 5.5 % per year (trend: $R_s = -0.0985y + 199.3$; $R^2 = 0.2829$).

SPATIAL AND TEMPORAL VARIABILITY OF SNOW THERMAL RESISTANCE

The patterns of thermal resistance depend on variations in the depth and density of snow. The R_s value increases during the first months of the cold season but the changes can be minor in the second half of winter in some areas [*Balobaev, 1991; Osokin et al., 2014*], because snow compaction can compensate for the R_s increase caused by increasing snow depth.

The climate conditions of the cold season and the parameters of snow in Russia were estimated using long-term annual means at weather stations north of 60° N, where route snow surveys were performed [*RIHMI–WDC*, *2013*]. Snow depth data, reported since 1966 at the cited website, were selected mainly

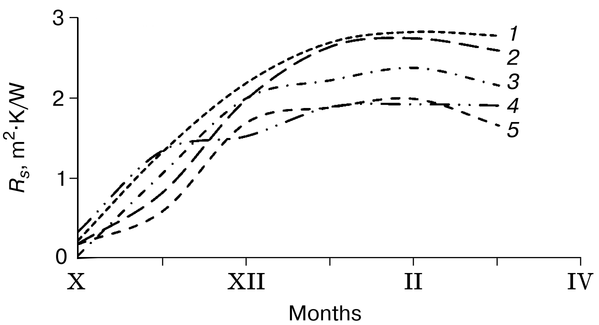

Fig. 2. Long-term means of snow thermal resistance (R_s) from October through March for the period 2001–2010 at different weather stations:

1 – Boguchany; *2* – Kolpashevo; *3* – Barguzin; *4* – Mogocha; *5* – Krasnoyarsk.

for stations at elevations within 300 m above sea level, as well as nine higher stations (eight at 400–700 m and one at 1315 m asl). Snow surveys were performed in the forest and in the field at 77 and 83 stations, respectively. After excluding 24 stations, where the surveys were performed in both forest and field landscapes, 112 stations were used eventually to study variations in the depth and density of snow.

The maps of snow depth and density in Russia and their long-term means for November, January and March for 2001–2010 published earlier [*Osokin and Sosnovskiy, 2014*] show that the March snow is the densest (0.30 g/cm^3) in the southern and northern areas of European Russia [*Osokin and Sosnovskiy, 2014*]. Snow becomes 10 % denser in March than in January; it reaches 0.25 g/cm^3 in West Siberia and only 0.18 and 0.16 g/cm^3 in Yakutia and Transbaikalia, respectively. The maximum snow depth in the end of the cold season (April–May) is 5–15 cm greater (on average) than in March.

Average March snow depths and densities were compared between the 2001–2010 and 1966–2000 datasets according to snow surveys at weather stations. These time spans were selected because annual air temperatures in 2000–2010 were anomalously warm in many areas [*Malkova et al., 2011*]. The current trends of warming rates were analyzed with reference to mean annual air temperatures for 1965–2000 and 2000–2010 reported by *Malkova et al.* [*2011*].

Long-term snow depth means for 2001–2010 are 40 % greater than those for 1966–2000 in the northern Tyumen region and in some areas of European Russia and West Siberia [*Osokin and Sosnovskiy, 2014*]; a smaller increase within 15 % is observed in northern Yakutia and east of the Lena River, as well as in central West Siberia. At the same time, the long-term snow density means decrease for 5 to 15 % in those areas and become 10 % greater in northern West Siberia in 2001–2010 relative to 1966–2000. The changes in the long-term means of snow depth and density for the two periods, which vary geographically over the territory of Russia, have created heterogeneous thermal conditions for permafrost.

The calculations of thermal resistance and effective thermal conductivity of snow by equation (1) neglect its temperature regime. The dynamics of thermal resistance, which is crucial for heat insulation capacity of snow, was assessed for each weather station according to long-term R_s means in March and January for the periods 1966–2000 and 2001–2010. Additionally, the long-term means of October through March over 2001–2010 were included for five stations.

Variations of long-term thermal resistance means of October through March 2001–2010 observed at some weather stations of southern Siberia (Fig. 2, Boguchany: WMO code 29282; 58°23′ N, 97°27′ E; Kolpashevo: WMO code 29231; 58°18′ N, 82°53′ E; Barguzin: WMO code 30636, 53°37′ N, 109°38′ E; Mogocha: WMO code 30673, 53°45′ N, 119°44′ E; and Krasnoyarsk: WMO code 29570, 56°02′ N, 92° 45′ E) show a greater R_s increase from October to December-January than in January–March (10 %).

The dynamics of R_s in the territory of Russia mapped as changes of long-term R_s means for January relative to those of March for the period 2001–2010 (Fig. 3, *a*) is as follows. The R_s difference between March and January does not exceed ±10 % over 80–90 % of Siberia, both in forest and field conditions. For instance, the color range 1.05–0.90 of the color scale in Fig. 3, *a* predominates nearly throughout Siberia, except for Transbaikalis and the Kolyma middle reaches, where the thermal resistance changes for January–March are similar to the pattern of Fig. 2. In central and western European Russia (where soils freeze up seasonally), the R_s value is 15–25 % higher in March than in January, while the relationship can be inverse in southern European Russia and southern West Siberia due to greater snow density in March. The January snow depth is 70 % greater than its March maximum over 75–80 % of the Siberian territory [*Osokin and Sosnovskiy, 2014*]. As a result of snow density increase [*RIHMI–WDC, 2013*], the R_s changes are most often small [*Balobaev, 1991*]. Thus, R_s can serve as a complex thermal parameter to characterize the heat insulation capacity of snow in specific areas.

The calculated long-term March R_s means for 2001–2010 (Fig. 3, *b*), according to snow surveys in the forest and in the field, vary in the ranges 0.7–5.0 and 0.3–3.3 $m^2 \cdot K/W$, respectively, the difference being due to greater depth and lower density of snow in the forest than in the field. The R_s values are the highest in the middle and lower reaches of the Yenisei and in northern European Russia in the forest snow data and in the Yenisei and Lena middle and upper reaches, as well as in central European Russia, according to the field snow surveys.

The changes of long-term means of snow depth and density for the period 2001–2010 relative to those of 1966–2000 cause respective R_s changes (Fig. 3, *c*). The long-term means of March R_s in 2001–2010 are 10–40 % larger than in 1966–2000 in many areas: southern, western, and northeastern European Russia; Komi Republic; northern West Siberia; southern Siberia and Russian Far East. Smaller R_s variations within ±5 % are restricted to central European Russia, northwestern Yakutia, and northeastern Krasnoyarsk region. In some areas, R_s become 5–15 % lower: central, eastern, and northeastern Yakutia (Lena-Kolyma interfluve); middle and lower reaches of the Vilyui; Chukchi district; middle reaches of the Yenisei; southeastern Tyumen region; and central Krasnoyarsk region.

THERMAL STABILITY OF PERMAFROST

The heat insulation capacity of snow expressed via its thermal resistance can compensate the climate impacts on the temperature and thermal stability of permafrost (see above). Possible changes to the permafrost thermal stability can be assessed by comparing changes in the long-term means of snow thermal resistance for March in the period 2001–2010 relative to those for 1966–2000 (Fig. 3, *c*) with the changes in long-term means of freezing index for the respective periods (Fig. 3, *d*). The absolute freezing index becomes 30 % lower (the greatest reduction) in southern European Russia; 7–15 % lower in West Siberia, and in western and central European Russia; 5–7 % lower in the Amur region, northern West Siberia, and southeastern Yakutia. The reduction does not exceed 5 % over 80–90 % of the Siberian territory. Namely, it is from 2 to 5 % in eastern and southeastern Yakutia, in the Chukchi autonomous district, in the southeastern Tyumen region, in the eastern Krasnoyarsk region, and in Transbaikalia; minor changes of ±2 % are observed in the upper and middle reaches of the Yenisei, Lena, and Vilyui rivers, as well as in the Yana catchment. Note that R_s changes more strongly in the 2001–2010 relative to 1966–2000 than the freezing index: 0.85–1.65 u.f. (see color scale in Fig. 3, *c*) against 0.70–1.02 u.f. (Fig. 3, *d*), or 2.5 times less.

The comparison of the long-term R_s and freezing index data for the two periods 2001–2010 and 1966–2000 (Fig. 3, *c*, *d*) shows that the minor (0–5 %) reduction in the freezing index in eastern and southeastern Yakutia, in the middle and lower reaches of the Vilyui, in the Chukchi district, in the southeastern Tyumen region, and in the central Krasnoyarsk region is compensated by a 10–15 % reduction in the thermal resistance of snow. Therefore, the climate changes in these regions are favorable for maintaining the thermal stability of permafrost. The conservation of permafrost and increase in its thermal stability are due to small changes in air temperatures of the cold season, as well as to reduction in the thermal resistance of snow, which accelerates freezing of the active layer (seasonal thaw) and cools down the shallow permafrost.

The conditions for permafrost conservation are the worst in southeastern European Russia, in southern Siberia, and in northern West Siberia, where the thermal resistance of snow increases (Fig. 3, *c*). This inference generally agrees with the results of *Malkova et al.* [*2011*]. They report that the conditions in the Komi Republic, in the middle reaches of the Yenisei, and in the Baikal region are unfavorable for permafrost stability which shows trends of rapid warming, poor stability, and degradation in zones of sporadic and high-temperature permafrost.

The thaw depth trends for the 1999–2008 decade are evident in the map of CALM sites (Circum-

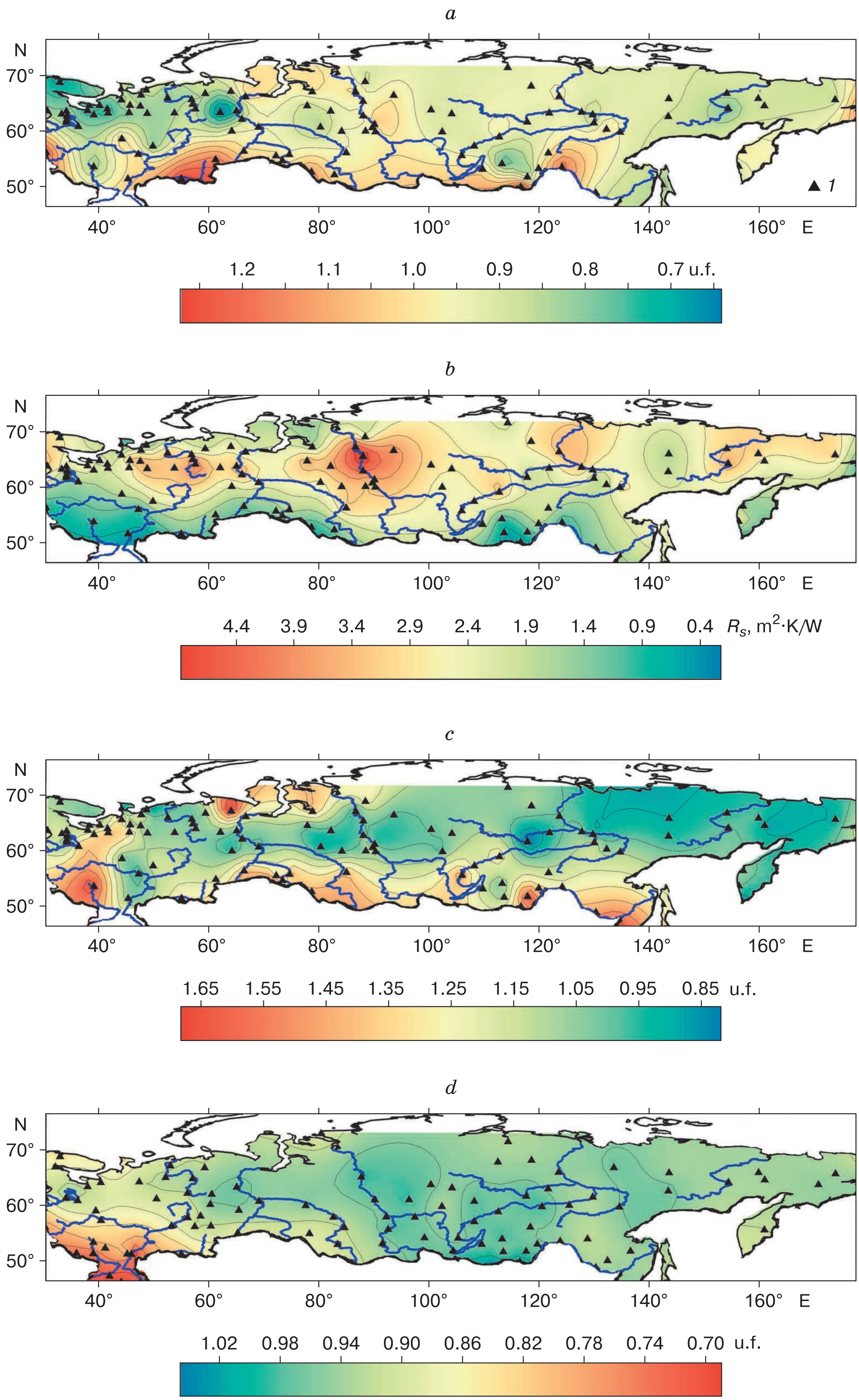

Fig. 3. Spatial and temporal variations in snow thermal resistance (*a–c*) and freezing index (*d*):

a: difference of long-term R_s means for January relative to March in the period 2001–2010; *b*: long-term R_s means for March in the period 2001–2010; *c*: ratio of long-term R_s means for March in 2001–2010 to the respective values in 1966–2000; *d*: ratio of long-term freezing index means for 2001–2010 to the respective values in 1966–200; triangles mark (*1*) locations of weather stations.

polar Active Layer Monitoring project) [*Anisimov et al., 2012*]. The sites with small (0–1 cm/yr) positive and negative trends are most numerous in Yakutia while the largest changes (more than 2 cm/yr) fall into the Komi Republic, where the freezing index changes insignificantly (Fig. 3, *d*). This behavior is consistent with the changes of snow thermal resistance which decreases for 10–15 % in eastern Yakutia but increases for 20 % in Komi (Fig. 3, *c*).

CONCLUSIONS

The thermal resistance R_s is a complex parameter that characterizes the heat insulation capacity of snow. Its change in the second half of the cold season does not exceed 10 % over most of the territory of Siberia (80–90 %). The reduced thermal resistance of snow for the period 2001–2010 at the Verkhoyansk weather station, cited as an example, has compensated the effect of decreased freezing index on the ground temperature, and even led to ground cooling at the depth 320 cm.

The long-term means of March snow thermal resistance R_s for the period 2001–2010 became 10–40 % higher than in 1966 through 2000 in many areas of Russia: southern, western, and northeastern European Russia; Komi Republic; northern West Siberia; southern Siberia and Russian Far East. This creates unfavorable conditions for ground cooling and freezing, which disturbs the active layer, increasing the seasonal thaw depth and reducing the seasonal frost depth. Thus, the active layer thickness can change due to warming, as well as to increasing thermal resistance of snow which impedes ground freezing. All these areas are unfavorable for permafrost conservation and thermal stability.

There are also areas where R_s reduction is smaller (5–15 %): central and eastern Yakutia; middle and lower reaches of the Vilyui; the Chukchi district; the southeastern Tyumen region; and the central Krasnoyarsk region. Minor changes in the cold season temperatures and reduced thermal resistance of snow in these areas are favorable for the thermal stability of permafrost.

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