Temperature Regime of Permafrost under Different Geographical and Geological Conditions

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Abstract

1. The temperature regime of permafrost depends on its composition and on the boundary conditions at the upper and lower surface of rock masses determined in turn by a complex of geological and geographical processes.

2. Examples of geothermal sections of rock masses are given by characterizing the different regime of temperatures which are explained by the peculiarities of composition and by historical and spatial variations of the upper and lower boundary conditions in different physico-geographical and geological circumstances (variation of climate, advance and retreat of the sea, glaciation development, tectonic mobility of territories etc.).

Permafrost is the upper layer of the lithosphere in which the temperature is continuously lower than 0°C irrespective of the content and phase state of moisture contained therein.

Thus it may be said that temperature is the main condition of the existence of permafrost. Therefore it is considered to be the most important characteristic indicating the intensity of the cryogenic process.

Temperature wise, permafrost is divided into two layers: the upper layer, reaching a thickness of 20–30 m where seasonal oscillations of the air temperature occur on the surface and which is called the thermoactive layer. The lower one-usually much thicker, where the temperature remains practically constant as a rule for a year or a number of years and varies by the depth in accord with the geothermal gradient.

The temperature at the bottom of the thermoactive layer is determined by heat exchange on the surface and is expressed by the general dependence principle (Principles of geocryology, 1959):

$$t_{TA} = t_b + \frac{A_{cm}}{2}\left(1 - \frac{1}{f}\right),$$

where t_{TA} is the temperature on the bottom surface of the thermoactive layer,
t_b the mean annual air temperature,
A_{cm} the annual amplitude of mean monthly air temperatures,

and

$$f = e^{+z\sqrt{\frac{\pi}{kT}}},$$

where z is the thickness of the snow cover,

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k the thermal conductivity of snow,

T the period of oscillations equal to one year.

In actual cases the influence of local factors should be taken into account. For instance the lithological composition, humidity (ice content) of rocks, hydrogeological peculiarities of plant and ground cover, channels and reservoirs, geomorphological processes (sedimentation, denudation, epeirogenic motions of the Earth's crust), human activity etc. So far these factors have not been studied quantitatively.

The temperature field of the lower layer depends both on the temperature at the bottom surface of thermoactive layer, which was mentioned above, and on the composition, structure and thermophysical properties of the lower layer and unfrozen rocks underlying it. Hydrogeological and geochemical processes can play a great role here as well as periodical, especially longperiodical (from dozens of years to hundreds of thousands of years) oscillations of the air temperature on the ground surface which penetrate to the layer of frozen rocks and deeper in connection with historical variations of climate. All these factors determine the value and constancy of the geothermal gradient in the lower layer of permafrost.

As a definite geographical event, frozen rocks of the Earth's crust are subjected to the laws of latitudinal and vertical zonality.

Temperatures at the bottom surface of the thermoactive layer, *i.e.* at a depth of 15–30 m from the surface, decreases in general from South to North ranging from 0°C close to the southern boundary of the region of distribution of permafrost to −13°C and lower—in the extreme North. However at many definite points with special boundary of frozen layers, rather sharp anomalies of regional character are observed and therefore break the relatively simple scheme of temperature changes of frozen layers shown on geographical maps.

Table 1 gives examples of temperature changes in frozen layers depending on different physico-geographical and geological conditions. These examples are taken from the districts of northern Asia and America within the limits of 62–75°N known by the intensity of freezing.

Considerable contrasts in the development of frozen rocks—their temperatures and freezing depth—are observed in mountainous regions. Thus for instance the role of the relief is distinctly observed in an example of Chukotka. Elevated elements of the relief are characterized by an extended geothermal gradient as compared with lowlands owing to lateral cooling from the side of the slopes. Temperature of frozen rocks in the uplands is higher than in lowlands which is connected with lower seasonal ice content. The thawing layer in the mountains is composed of coarse waste material. In the Chukotka lowlands this layer is formed in the highly wet summer and because of its loamy peat formation it containes much ice in winter.

Quite unexpected results of temperature measurements of frozen rocks were obtained in the region of the so-called pole of coldness in the northern hemisphere at Oymyakon. The pole of coldness of mountainous rocks does not coincide in geographical respect here with the pole of coldness of air temperature. In Oymyakon there is a depression with a considerable low mean annual air temperature and poor snow cover. Here the pebble-beds filling the depression appear to be more heated than the argillites in the

Table 1. Temperature (T) of frozen rock masses on the bottom surface of the thermoactive layer (TA) and geothermal phase (g) in different physico-geographical and geological conditions

Region of observations	Latitude of the site	Absolute height	Mean annual air temperature (°C)	Position in the relief	Geological structure	T on the bottom surface of the TA layer (°C)	Depth of occurrence of geoisotherm	Mean gradient (m/°C)	Reference
1	2	3	4	5	6	7	8	9	10
Chukotka, the Anadyr River mouth	65°	6	− 7.0	a) Primorskaya (sea-side) lowland	Icy loamy sandy pebble-beds of the Quaternary rocks	− 5.2	200	40	Nekrasov, 1960
				b) Mountainous ridge	Basalts	− 3.2	250	73	Nekrasov, 1960
Yakutia north-eastern mountainous part	63–62°	740	−16.5	a) Oymyakonskaya depression, river valley	Sandy pebble-beds. Quanternary sediments	− 7.2	600 ?	—	Grave *et al.* 1964
		2 063	−14.7	b) Suntar-Khayat ridge, saddle of glacial valley	Rock waste and fractured argillites	− 9.1	—	—	Grave *et al.* 1964
North-East of the USSR, the basin of the upper reaches of the Kolyma River	63°	779	−12.4	a) Valley slope 300 km from the river-bed	Keratinization-like sandstones and shales	− 3.8	263	55	Kalabin 1960
		603	− 9.4	b) River valley	Productive sandy-argillaceous suite with brown coals	− 2.7	81	14	Kalabin 1960
Mid-Siberian Plateau	66°	500	−12.5	a) River valley	Dolomites, limestones of lower P_2 and PCm; at the depth of 1 830 m horizon of supercooled mineralized waters	—	1 500	130	Melnikov 1965
		500	−12.5	b) River valley	Carbonaceous rocks, pressure mineralized waters	− 5.0	500	100 ?	Melnikov 1962
Alaska Point Barrow	71°	15–30	− 5.5	Sea-side lowland					
				a) near the sea	Ice, sea sands and clays on sandstones	− 7.3	200	23.4	Lachenbruch *et al.* 1962
				b) far from the sea	Ice, sea sands and clays on sandstones	− 9.8	400	40.1	Lachenbruch *et al.* 1962
Arctic Canada Cornwallis Island	75°	10	−15.0	Sea shore 400 m inland from the sea shore	Dolomites	−13.5	400	25.4	Lachenbruch 1957
				Calculated by assuming that the region had been never covered by sea		−14.0	1 000	43.0	Lachenbruch 1957

high mountainous Suntar-Khayat ridge located near-by and which is characterized by its remarkably softer climatic regime. It was surmised that a certain role was evidently played by hydrogeological conditions of intermountainous depressions which are collectors of underground waters, flowing down from the surrounding ridges and preventing cooling of mountainous rocks and also having a low content of the seasonal thawing layer.

Another example of hydrogeological and an equal extent of geochemical influence on the intensity of freezing of mountainous rocks on Mid-Siberian Plateaus, near the polar circle, is of interest.

Here in the Markhi River valley and at the site of its inflow, *i.e.* on the Daaldyn River (the Viliuy River basin), the deepest freezing of rock masses in the world, namely, the freezing of rock mass more than 1 000 m thick with comparatively high mean temperature in the order of $-5°C$, was displayed by deep bore holes. The large extension of geothermal gradient is apparently connected with the cooling effect of strongly mineralized supercooled underground waters flowing down from the Anabara anticlinorium and observed under the layer of properly frozen rock masses.

As another example in another district in the basin of the upper reaches of the Kolyma River we observe a geochemical heating effect: an exothermal reaction occurring in coal bearing rocks associated with oxidation of sulphides contained in them sharply limits the freezing depth (to 80 m) irrespective of severe climatic conditions. Beyond the sphere of this exothermal effect the rocks have a more extended geothermal gradient and temperature of 0°C was observed a depth of more than 250 m.

Both the value of temperature and its change in depth are essentially influenced by basins including seas, lakes and rivers. This influence is illustrated by an example taken in Point Barrow, Alaska. Near the sea and the fresh water lake frozen rock masses composing the sea-side lowland have higher temperatures and the temperature distribution by depth is characterized by a relatively short geothermal gradient. In contrast to the above, a lower temperature and a deeper occurrence of lower surface of frozen rock masses is observed far from the sea shore and the lake.

The influence of change of the conditions on the surface owing to historical reasons in temperature and thickness of frozen layers have been hardly studied so far. The most precise studies carried out by Lachenbruch (1957) in Arctic Canada (Cornwallis Island) on the shore of Resolute Bay allowed the statement which says that the temperature of frozen rock masses and geothermal gradient are in certain discordance with the climatic conditions of this region. The above-mentioned researcher proved that the current thermal state of frozen layers at that site reflects a relatively recent time of the beginning of freezing of the Earth's surface, namely, 7 500 years have elapsed after a rapid retreat of the sea covering the territory of this district at a previous time. In other words, if the retreat took place earlier, the depth of freezing on Cornwallis Island would have been much deeper.

A curious example of a complicated historical effect on freezing of the Earth's crust was discovered in Western Siberia not so long ago. In the North permafrost reaching a thickness of 500–600 m and having a temperature of up to $-8°C$ has, as paleogeographic studies and mathematical calculations showed, a continuous area development which occurred mainly after the advance of the sea was found.

The southern permafrost is of a bi-strata structure: the upper layer of 30–80 m is underlain by thawed rocks and at the depth of 100–150 m the second frozen layer occurs. The thickness reaches 200 m. Still southwards (up to 61°N) frozen rocks occur only at a considerable depth from the surface (within a depth range from 100–200 to 300–400 m). Deeply occurring permafrost is a relic of a glacial epoch and it does not occur in the North which was covered by the sea at that time.

The thawing interlayer in the profiles with bi-strata structure of permafrost is connected with the maximal short-lived advance of the sea within the country (Baulin, 1962).

Thus complications and mixed characters of physico-geographical and geological conditions of the region of the Earth's crust permafrost, gave birth to essential anomalies in its zonal development.

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