Climate Change Research: Evaluation and Policy Implications
S. Zwerver, R.S.A.R. van Rompaey, M.T.J. Kok and M.M. Berk, (Eds.)

The response of permafrost ecosystems to climate change

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Abstract
Permafrost areas are extremely sensitive to change as permanently frozen ground is directly temperature dependent. The heat exchange interactions of climate and permafrost in the so-called "buffer layer" are highly complicated, as ground temperatures are strongly influenced by local factors (snow cover thickness and duration, vegetation, organic layer and soil characteristics), which are interrelated with climate. Variations in these variables may either enhance or counteract each other, which makes it difficult to predict the accumulated effect of all changes. However, several simulation experiments indicate large shifts of permafrost boundaries due to a temperature increase, resulting in extensive permafrost degradation (thermokarst). Geothermal profiles of the upper 100-200 metres of permafrost, which yield a temporally integrated record of air temperature changes in the past decades to centuries, show significant changes. However, the quantitative relationships between permafrost degradation and biogeochemical processes, including the generation or uptake of carbon dioxide and methane are still largely unknown.

1. INTRODUCTION

Approximately 25 % of the land surface of the Northern Hemisphere is underlain by permafrost. A major part of this huge area is designated as discontinuous permafrost (approx. 17.3 million square kms), the southern boundary of which roughly coincides with a mean annual air temperature of -1 to -2°C. Near its southern boundary it occurs in isolated patches or islands and is sometimes referred to as sporadic permafrost. Approximately north of the -6 to -8°C isotherm continuous permafrost (approx. 7.6 million square kms) occurs. Moreover, an area of approximately 2.3 million square kms, mainly at lower latitudes, is covered by Alpine or mountain permafrost. Permafrost areas will be among the most heavily affected parts of the world in the event of accelerated future warming [1, 2, 3, 4]. The objectives of this review paper are: 1) to emphasize the complex interrelations in the atmosphere-"buffer layer"-permafrost system, 2) to summarize permafrost response to past and future temperature changes and 3) to indicate the uncertainties with respect to permafrost ecosystems as sources or sinks of carbon dioxide and methane.

2. HEAT EXCHANGE AND THE ACTIVE LAYER

In permafrost areas several types of temperatures are defined (Fig.1), depending on where they are measured [1, 2]. The mean annual air temperature (MAAT) usually is several degrees lower than the mean annual ground temperature (MAGT), the latter being defined as the ground temperature at a depth where temperature fluctuates by less than 0.1°C per year. Above this depth the ground is subjected to strong seasonal fluctuations. Nevertheless, mean annual ground surface temperature (MAGST) can be deduced by upward extrapolation of the geothermal gradient, provided the measured gradient has achieved equilibrium and there are no recent climatic changes. Extrapolation of the geothermal gradient downwards will lead to an approximation of the depth of the permafrost base. Where the geothermal heat flow is constant, the geothermal gradient is inversely proportional to conductivity. The thermal conductivity in its turn strongly varies depending on soil properties and sediment texture. The water or ice content is especially critical. The geothermal gradient in different types of sediment ranges from about 1°C/60m for sandy, relatively ice-rich material (high conductivity ~ low gradient ~ thick permafrost) to about 1°C/22m for fine-grained, relatively ice-poor material (low conductivity ~ high gradient ~ thin permafrost).

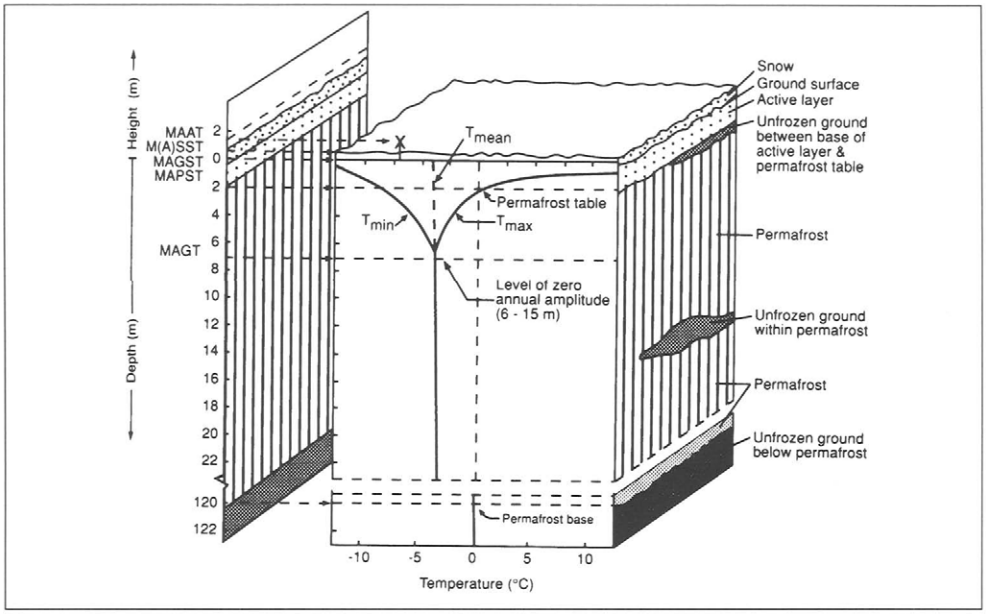

Figure 1. Permafrost terminology and ground temperature profile.

The heat exchange interactions of climate and permafrost through the active layer are strongly influenced by the vegetation cover, the (seasonal) snow cover thickness and duration, the organic soil (if present) and the mineral soil. Variations in these variables may either enhance or counteract each other [1, 2]. The snow cover thickness and duration probably is the most important factor due to its insulating properties. The conductive capacity of dry new snow is 1/5 of that of old compacted snow or dry sand, or 1/20 of that of wet sand or 1/25 of that of ice. Generally, a snow cover keeps the ground warmer, as it can heat up in summer, but is hampered in cooling down in winter. Snow cover thickness and duration depend to a large extent on the presence and nature of vegetation. Vegetation mainly has an effect upon surface temperatures by shading, thus cooling the ground. Moreover, vegetation prevents radiation back into the sky at night, and the soil is dried by evapotranspiration and this decreases its heat capacity. The organic layer strongly influences ground temperatures by its differential insulation capacities. In summer conductivity of a dry peat layer is extremely low (< 0.1 WmK). In winter, however, the organic layer freezes leading to conductivities many times higher (> 1.0 WmK), and consequently the ground cools off. To a lesser extent the properties of the mineral soil also influence the conductivity. Both changes in snow regime and in vegetation will determine the moisture condition of the soil and thereby the thermal conductivity of the materials.

3. HISTORIC CHANGES IN GEOTHERMAL REGIME

The above-mentioned uncertainties notwithstanding, in principle a rise in MAAT and consequently in MAGST will have the following effects. Firstly, the thickness of the active layer will increase. Secondly, the temperature profile within the permafrost will adjust itself to the new MAGST. The rate with which this happens depends on the thermal conductivity of the permafrost and the ice content of the ground. Response times of the active layer are in the order of years to tens of years. Eventually permafrost will decrease in thickness.
During Pleistocene glacial episodes the permafrost area was probably twice as extensive as the present-day extent, whereas during the "climatic optimum" of the Holocene the southern limit of discontinuous permafrost in the Soviet Arctic was up to 600km north of its present position. In historic times significant changes in permafrost zonation have also been documented; e.g. in the southern part of the discontinuous permafrost zone in Manitoba (Canada) the southern limit has shifted northwards over the past 150-200 years and the areal extent of permafrost terrain has diminished strongly. In the Mackenzie Valley (Arctic Canada) MAGT values increased by about 3°C during a recent warm period (late 1800s to the 1940s) and have since decreased about 1°C, resulting in a shift in the continuous-discontinuous permafrost boundary of several hundreds of kms.
Thus, the analysis of permafrost temperature as a function of depth appears to yield an integrated record of air temperature changes in the past. This has been well-documented by temperature profiles obtained from boreholes in the Alaskan Arctic Coastal Plain [5]. A vast number of these temperature profiles shows a distinct curvature towards higher temperature near the surface. The exact onset of warming seems to vary between locations, but they all indicate a warming in the range of 1.5-3°C during the last century, which seems to be in agreement with a similar trend in air temperatures as shown by regional weather records. Time series of annual permafrost ground temperatures in shallow drill holes (<60m) in northern Alaska show that these temperatures have cycled in the decade from 1983 to 1993. Whatever the causes for this cyclic event may be (changes in air temperature, snowfall, solar irradiance), this detailed record clearly illustrates the rapid response of permafrost ground temperatures to external factors [6].

4. SIMULATING PERMAFROST ZONATION CHANGES

The greatest changes can be expected to occur in those parts of the discontinuous permafrost zone where MAGT values are close to 0°C. A model of heat and water transport has been constructed to simulate changes in the extent of permafrost in the USSR [7]. The model results based on a mean 2°C global warming (assumed to represent a 7-9°C increase in winter and a 4-6°C increase in summer in the latitudinal zone 55-70°N) suggest that in about 50 years the area occupied by continuous permafrost might be 15-20 % smaller than at present. The vertical movement of the permafrost table is expected to be 0.5-0.8m/yr during the first year of the model run and to decrease exponentially over time. After 50 years of the model run the maximum depth of the thawed layer was 7m. Comparable rapid changes in active layer thickness have been computed by others [8]. All model simulations suggest that due to climate warming the active layer in continuous permafrost will thicken, but that permafrost will remain, in contrast to discontinuous permafrost which might locally disappear completely [9].

5. PERMAFROST ECOSYSTEMS

The cold climate, short growing season and nutrient-poor soils in northern ecosystems have led to adaptations in plants and in soil organisms. In addition, the low rate of decomposition relative to primary production has caused a large accumulation of carbon as dead organic matter during the Holocene [10]. According to various sources northern (tundra and taiga) ecosystems contain 400-550Gt of carbon stored as soil organic matter [2]. Present-day emission of methane is dominated by northern wetlands, especially between 50-70°N, and particularly in western Siberia and east-central Canada where extensive bog and fen areas occur. Wet coastal arctic tundras could become a net carbon dioxide source at higher temperatures. Even slight decreases in water level due to higher evapotranspiration could strongly accentuate the temperature effect. This effect seems to be stronger than the direct effect of increasing atmospheric carbon dioxide content on carbon uptake in the tundra. Net primary production will increase somewhat due to increasing carbon dioxide levels, provided that nutrients are not limiting growth. However, in periglacial environments nutrients usually are a limiting factor. On the other hand, if temperature rises and drier conditions prevail, decomposition of organic matter may accelerate, causing more nutrients to become available. Warming of Arctic and boreal wetland ecosystems will almost certainly be followed by increased methane emissions, which currently account for the release of about 25-40Mt of methane to the atmosphere annually. It is estimated that a 4°C rise in these regions could lead to a 45-65% increase in methane release. If warming is accompanied by drying, then there may ultimately be a reduction of methane release. The only valid conclusion at present can be that the net results of these interactive effects are poorly known [11, 3].

6. CONCLUDING REMARKS

Widespread thermokarst development is likely to occur as a consequence of permafrost degradation, particularly of ice-rich ground [7]. Regional lowering of the permafrost table will cause the development of large unfrozen near-surface aquifers with perennial groundwater flow. In practical terms, increased terrain instability, especially in the first phases of permafrost degradation, would lead to major concerns for the integrity and stability of roads, railways, airstrips, dams, reservoirs and other foundations in affected areas. Deepening of the active layer would subject foundations to continuing deformations as a result of thaw settlement. Many of the technologic and socio-economic implications of climate change at higher latitudes were recently reviewed [12, 13]. However, there is an acute need for a circumpolar network of ground temperature stations, ideally at or near ITEX (international Tundra Experiment) locations, to provide information about the nature, extent and rate of permafrost change that will be necessary for realistic regional and global planning activities [3, 12].

7. REFERENCES

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