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Thermic and humidity relations of chosen Spitsbergen soils during spring ablation of the snow cover

ABSTRACT: Studies on the thermics, humidity and refreezing rate of two types of periglacial soils of the marine lowlands of the Hornsund area (SW Spitsbergen) were carried out during spring ablation of the snow cover (1.06.—30.06.1987). Structural soil and peat soil (moss community) were chosen. The soils were studied in places almost completely devoid of the snow cover. A considerable differentiation of temperature in vertical profile and of humidity of surface layer in both soils were found. The dynamics of ground water table and the course and depth of spring refreezing of both soils were also different. The differences reflect the different origin of soils and consequently, their different grain-size distribution, physical properties and morphology of both types of soils as well as their water balance.

Key words: Arctic, Spitsbergen, periglacial soils, temperature and humidity.

Introduction

In June 1987 in the Fulgeberget drainage area (Hornsund, Spitsbergen) investigations were carried out which aimed at the understanding of the thermic and humidity conditions of two types of periglacial soils during spring ablation of snow cover. Observations were a part of parallel ecological studies aiming at learning of the intensity of metabolic processes in soils of Spitsbergen tundra (Bieńkowski 1990, Fischer 1990).

Two soil types were chosen for study: a structural soil with an initial stage of stony tundra assemblage, and a peat soil (moss community — *Calliergon stramineum* — *Sanonia uncinata* assembly; Grodzińska and Godzik, unpubl. data). Both stations were situated in marine lowlands (low marine terraces were at the altitude of 10 and 25 m a.s.l.; Krawczyk and Pulina 1982, Pulina et al.

1984). The soils were studied in places almost completely devoid of the snow cover. In this way the effect of the diversified snow cover ablation on thermic relations, soil humidity and the rate of soil refreezing was excluded.

The observations aimed at the determination of thermic and humidity diversity of both types of soils with an attempt to explain the causes of these differences. Assuming the generally identical meteorological, geological and geomorphological conditions in the whole investigated area also an attempt was undertaken to determine the influence of water conditions and vegetation upon the magitude and refreezing rate in the soils observed.

Investigated area and methods

Investigations were carried out at two stations where measurement instruments were installed (Fig. 1). Air temperature at 5 cm above the ground was measured with Assman psychrometer. The ground temperature at depths of 0, 5, 10 and 15 cm was measured with a mercurial soil thermometer at fixed time: 6, 12 and 18 GMT. Additionally maximal and minimal temperatures of the ground surface were measured in the terms mentioned above. The movement of ground water table was measured using a weekly limnigraph installed on piezometers. Thickness of the refrozed ground layer was determined with an accuracy of 1 cm by probing with a steel rod; the measurements were performed 3 times in 24 hour period. The humidity of the soil was determined by dryer method and samples for analysis were taken once per 24 hours in the morning.

The station where the peat soil was studied was located under Ariekammen slope within the elevated marine terrace situated at 28 m a.s.l. The measurement instruments were located at 28 m a.s.l., on one of abundant solifluction lobes (Jahn 1970) of the surface inclination of 5°40′. It was covered by the peat soil layer 10 cm thick. In vertical profile of this soil one can distinguish three main layers: a layer of 0—1 cm was a living moss (*Calliergon stramineum*), a layer from 1 to 6 cm was dead but not decomposed moss, whereas the layer from 6 to 10 cm was a slightly decomposed moss brown-green in colour. The full water capacity of this layer was 1065%. The moss community covered the gravel-type ground that can be called loamy gravel of the following grain size distribution: clay — 10,6%, silt — 6% and sand — 27%; the remaining 56,4% was fine gravel with a maximal diameter of 20,5 mm. During the spring ablation of snow cover the soil at this station was characterized by the erosion-lessivage type of water relations.

The station for study of structural soil wass situated on the plain of elevated marine terrace at the foot of Fulgeberget mountain at 12 m a.s.l. The plain is an extensive area with a slope of 1°30′, covered by weathered material where a polygonal net of mud craters and of stone circles occurred. According to Jahn

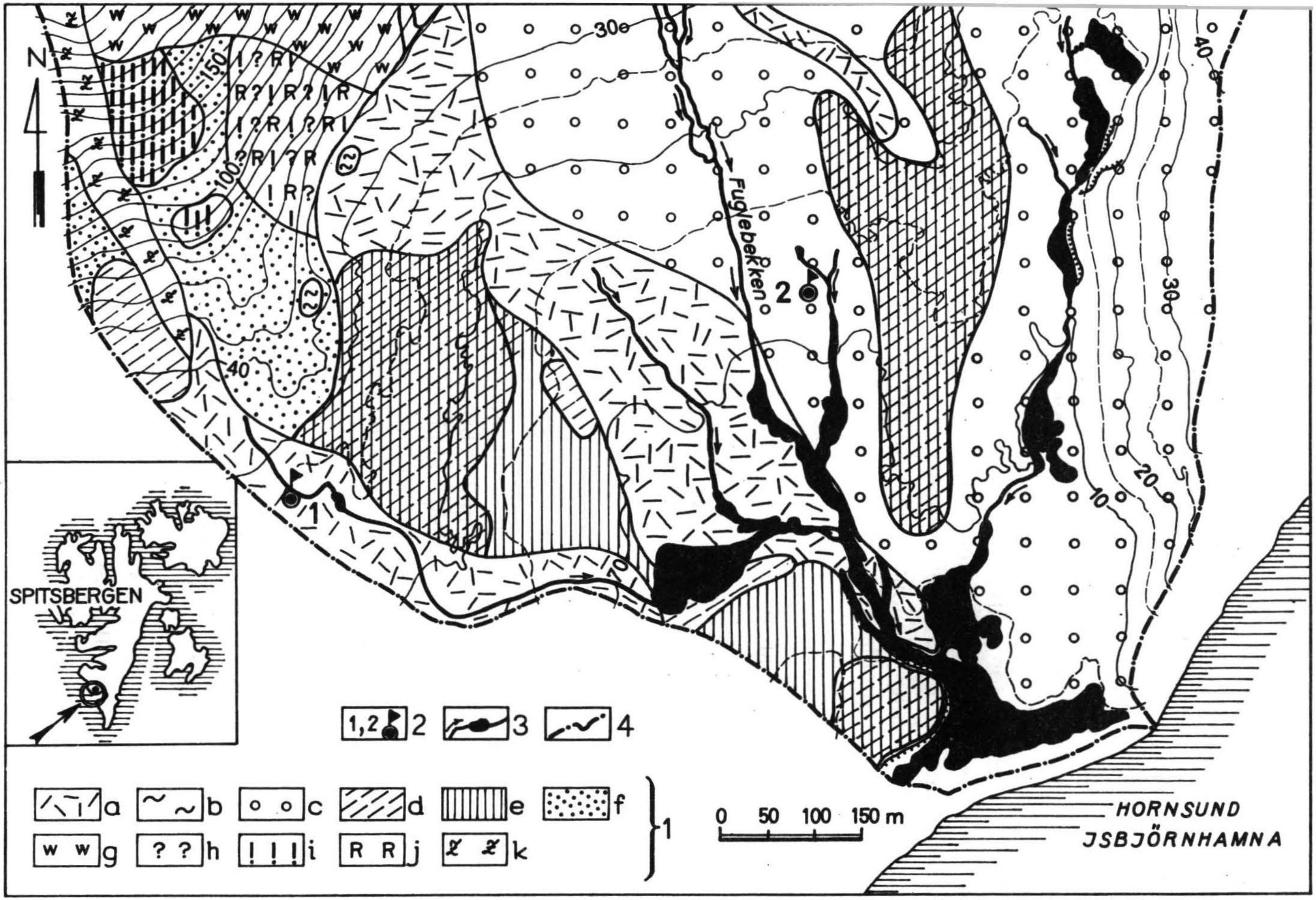

Fig. 1 Investigated stations in the lower part of the Fugleberget drainage area (Hornsund) and the vegetation assemblages;

1 — vegetation assemblages (according to Grodzińska and Godzik, unpubl. data): a — *Calliergon stramineum* — *Sanonia uncinata* b — *Tetraplodon mnioides* — *Aplodon wormskjoeldii* c — stony tundra d — *Cladonia mitis* — *Cetraria nivalis* — *Rhacomitrium languinosum* e — *Sphaerophorus globosus* f — *Chrysosplenium tetrandrum* — *Cochlearia officinalis* — *Cerestrium alpinum* g — *Sanonia uncinata* — *Aulaconium turgidum* — *Saxifraga oppositifolia* h — *Sanonia uncinata* i — *Candelariella arctica* j — *Xanthoria elegans* k — saxicolous lichens; 2 — stations: 1 — moss community, 2 — structural soil, 3 — water bodies, 4 — limits of the drainage area.

(1970) it belongs to the gravitation type of transformed grounds. The mesurement instruments were installed here in one of mud craters almost completely devoid of vegetation (the assemblage of stony tundra in initial stage). The grain-size distribution of the material building a crater (in the layer from 0 to 0,5 m) was following: the finest particles made 16,3%, silt — 18,2%, sand — 24,6%; the remaining 40,9% were gravels (maximal diameter of 18,5 mm). Full water capacity of this soil in the surface layer was 36,3%. During the spring snow cover ablation this soil had an evaporation — hydromorphic type of water relations.

The soils chosen for study had different origin and, consequently, different physical properties and morphology. The two soils studied were also differently situated within the Fulgeberget drainage area; this resulted in their different thermic and humidity relations because of different rate of exchange of heat and water in soil profiles.

Results and discussion

Thermic conditions in Hornsund region in summer

Average annual air temperature for Hornsund area for the period 1978—1986 was very close to many-years one and amounted to −5.0°C (Rodzik and Stepko 1985, Ustrnul 1987). The average annual air temperature varied between −2,3° and −6,2°C and average precipitation varied between 298 and 504 mm. In the annual course maximal temperature occurred in June (4,5°C) and minimal — in January and February (−11,9°C). It was confirmed also by the data of Baranowski (1968) and Grześ (1985). In June 1987 at Polish Polar Station in Hornsund the average monthly air temperature was 2,1°C (Fig. 2) and the average ground temperature at a depth 5 cm was 2,9°C.

According to the thermic criterion (Baranowski 1968) in SW Spitsbergen spring lasts on the average 26 days, from May 25th to June 19th (Ustrnul 1987; mean values from 1978—86). In 1987, in the Fugleberget drainage area winter ended on May 14th and spring lasted 44 days: from 15th of May to 27th of June. In early spring the average air temperature within 24 hours ranged between −1,9°C and +1,9°C. From the 7th of June to the beginning of summer it remained above 0°C (min. 1,1°C, max. 4,7°C). The fluctuations of average temperatures within 24 hours in spring reached maximum 6,6°C and the average air temperature in This period was 1,1°C.

In June the average air temperature within 24 hours at 5 cm above the ground was 3,7°C in the case of structural soil and 3,4°C in the case of peat soil. Average air temperatures within 24 hours above the structural soil during the whole month were a bit higher (by 0,1°C to 0,8°C) than above the moss community. Considerable differences were noted in temperature fluctuations within the 24 hour cycle. During warm days (average air temperature above 4°C) maximal temperature amplitude above the structural soil was 9°C, and above the moss community — only 7,4°C. During cold days (average air temperature below 2°C) maximal temperature fluctuations above both soils were similar, not exceeding 5°C. During the day, above the structural soil the maximal air temperature reached 13,0°C, whereas above the peat soil the maximal temperature was 10,6°C.

The thermics of studied soils

Heat input to the surface of both soils studied was similar, but different was its „management" by the soil. Different physical properties of soils and their different humidity in the whole studied period caused a different heat exchange between the atmosphere and the ground; heat accumulation in vertical profiles was also different. The course of average ground temperatures within 24 hours in both types of soil in June is presented in Fig. 2. During the whole month at all depths studied the temperature of structural soil was higher than that of the peat soil. At that time, at the ground surface, the highest average twenty-four hours soil temperatures were noted: at the peat soil surface — 10,2°C, at the structural soil surface — 12,0°C. Extreme temperatures at the ground surface considerably exceed air temperatures and their amplitude in Spitsbergen conditions can reach over 20°C (Kamiński 1985). The maximal temperatures of the ground surface noted in June were 21,4°C (structural soil) and 19,2°C (peat soil). Twenty-four hours average temperature at the surface of both soils studied during all June persisted above 0°C and the average monthly temperatures were: for peat soil surface — 5,5°C and for the surface of structural soil — 6,1°C.

At the depth of 5 cm (Fig. 2) the run of average twenty-four hours temperature in both soils had similar shape but was very different in temperature values. The differences between structural soil and peat soil at that depth ranged in June from 0,5° to 4,2°C. In the structural soil the temperatures noted were always higher than in the peat soil. Average monthly temperature of structural soil at this depth was by 2,2°C higher than that of the peat soil.

At the depths of 10 cm and 15 cm similar situation was observed (Fig. 2). However, the differences in average twenty-four hours temperatures were much clearer (the largest differences were noted in third decade of June and amounted to 3,8°C on the average). With its great heat capacity the peat soil is a cold soil. Positive temperatures at the depth of 10 cm appeared here 7 days later than in the structural soil. Average monthly temperature of the structural soil in June at the depth of 10 cm was by 2,3°C and at the depth of 15 cm — by 1,6°C higher than that of the peat soil.

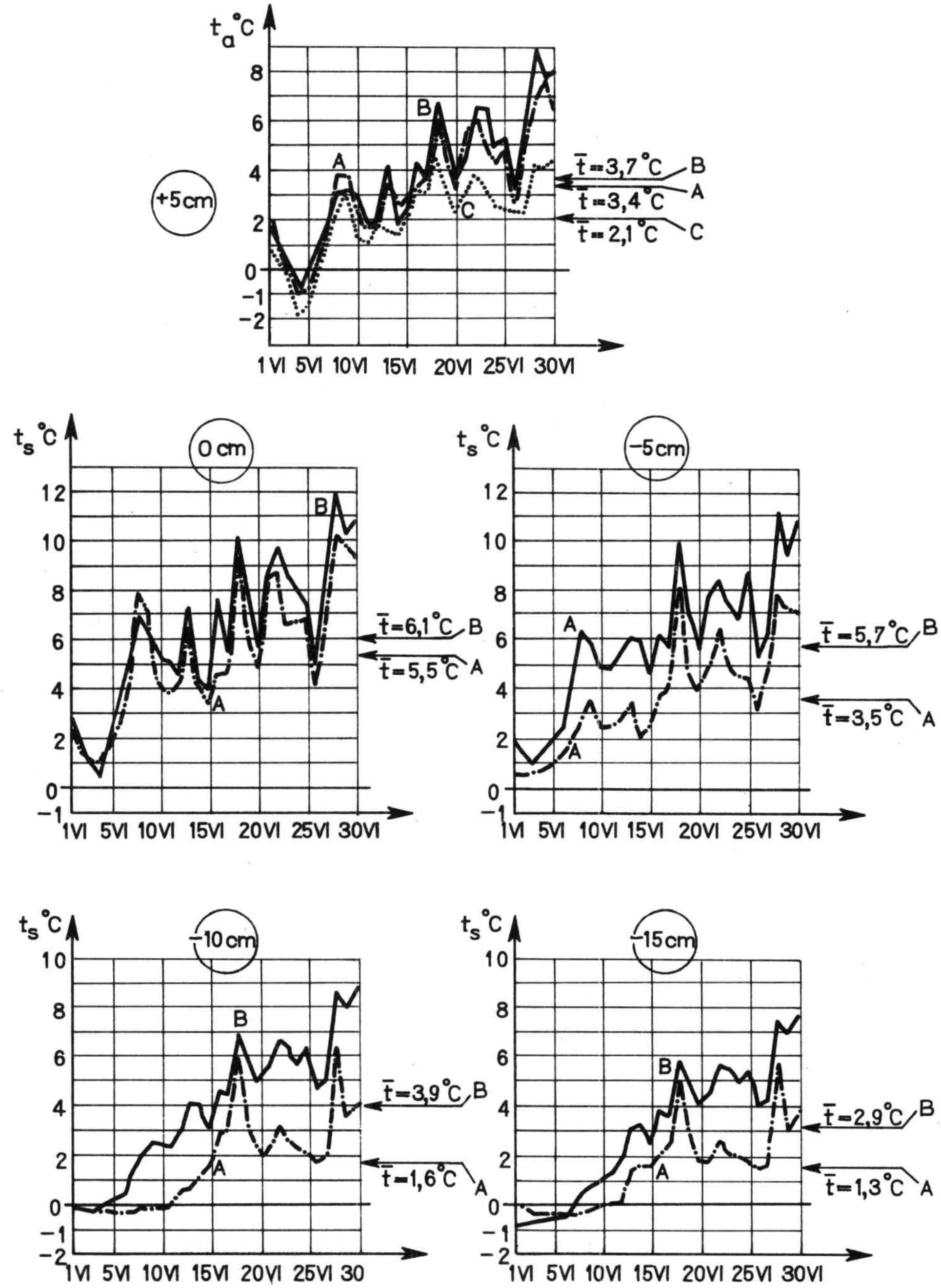

Fig. 2 Average twenty-four hours air temperature 5 cm above the ground and in the ground at the depths of 0, 5, 10 and 15 cm in two types of tundra soils in June 1987:

A — moss community (assemblage *Calliergon stramineum* — *Sanonia uncinata*)

B — structural soil (stony tundra assemblage)

C — air temperature at 2 m above the ground (meteorological data of Polish Polar Station).

The course and depth of spring refreezing of studied soils, ground water movement and the humidity of surface layer

The depth of seasonal refreezing of ground surface in Spitsbergen is discussed in many climatological and geomorphological papers (Czeppe 1960, 1966; Baranowski 1968, Jahn 1970, 1982; Grześ 1985, Migała 1988, Repelewska-Pękalowa, Gluza and Pękala 1988). Taking into account the air temperature course in Hornsund region, the period of snow cover disappearence and his own observations of marine lowlands Grześ (1985) assumed the half of June as an average period of the beginning of summer ground refreezing. This period proposed is by far a conventional one, since in its first stage the refreezing process is interrupted by short coolings and repeated freezing to the depth of dozen centimeters (Czeppe 1960). In glaciological literature as the beginning of ablation period usually the 1th of June is assumed (Baranowski 1977).

An essential factor affecting the course and the depth of ground refreezing is the precipitation in form of snow as well as rain. The presence of snow cover disturbes the development of active permafrost and a spatially differentiated snow ablation is a cause of unequal ground refreezing (Migała 1988). However, as it was mentioned above, the study stations were located in places almost completely devoid of snow. Thus, the effect of snow cover on the rate of refreezing of the soils studied could be neglected. The refreezing of soils is considerably accelerated by rainfalls which amount to 65% of annual precipitation (Grześ 1985). During the investigation period (June 1987) the precipitation amounted only to 1,5 mm. Therefore it was not a factor influencing soil refreezing.

The relationship between soil refreezing and its humidity is very complex. Generally it can be said that the humidity increase conduces the increase of heat conduction of the soil. Thus the refreezing velocity increases as well. The increase of heat conduction is di stinct only before the soil reaches a capillary water capacity. In the case of the structural soil the amount of water retained (in the whole observation period) was a stimulus of the rate of ground refreezing. In the case of peat soil which was situated on a way of surface run off of cold waters from melting snow — the water was an insulating factor inhibiting the development of active layer.

The heat-insulating role of organic layer (turf, peat) in the refreezing of the ground is commonly known. The decrease of annual ground temperature amplitude under the moss-peat layer by 50—60% can be observed (Kudrjavcev 1978). The studies by Czeppe (1966) and Jahn (1970, 1982) showed that the moss layer of the thickness of about 10 cm distinctly decreased the refreezing rate. It was noted, however (Repelewska-Pękalowa, Gluza and Pękala 1988), that the vegetation did not play an important insulating role when water was moving under the snow cover.

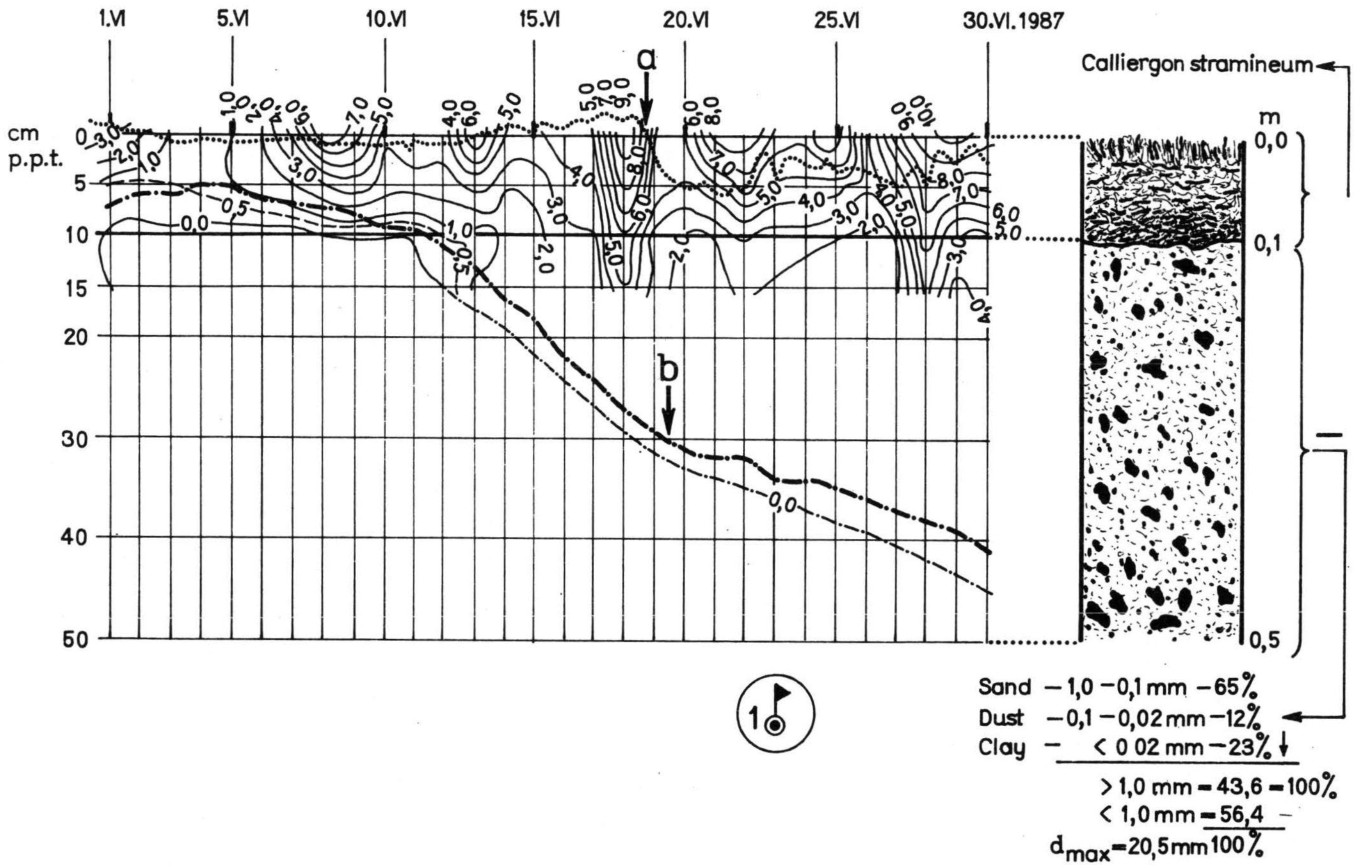

Fig. 3 Thermoisopleths of the surface layer and the dynamics of ground water table and permafrost during the snow cover ablation — moss community.

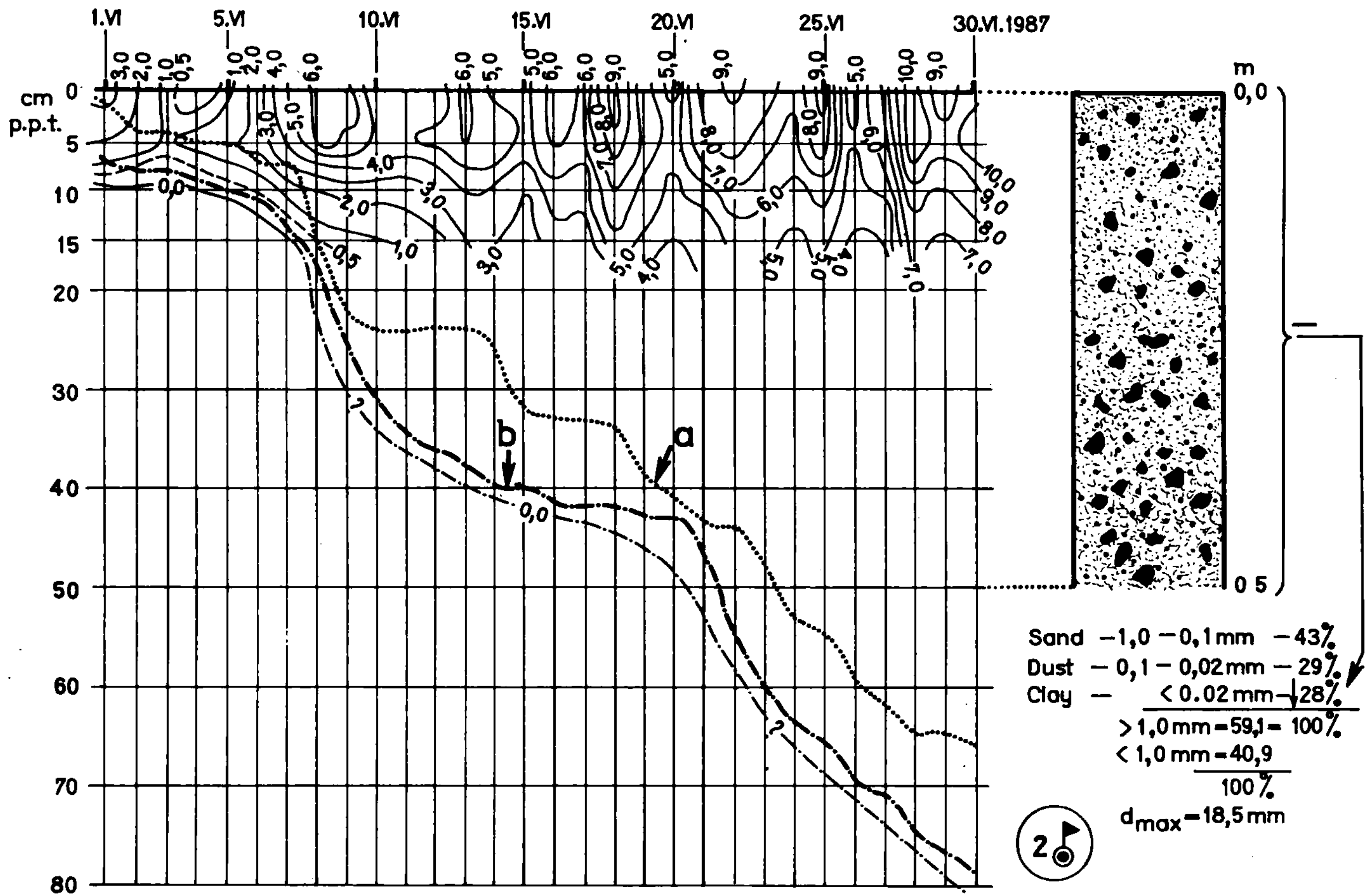

Fig. 4 Thermoisopleths of the surface layer and the dynamics of ground water table and permafrost during the snow cover ablation – structural soil.

The rate of the peat soil refreezing during the spring snow cover ablation was determined mainly by the fact that the station studied was situated on the way of run off of the waters originated from the melting snow cover (Fig. 3). Cold ablation waters were an insulating layer during the whole period of study, delaying the beginning and inhibiting the rate of soil refreezing. This delay, in relation to structural soil, amounted to 7 days. The average twenty-four hours rate of peat soil refreezing was 1,2 cm and did not exceed 2,1 cm. The refreezing proceeded uniformly in the whole period of the spring snow cover ablation and no inhibition or acceleration of this rhytm has been observed. In June, the permafrost level has lowered by 35 cm and at the end of polar spring it reached the depth of 41 cm. The dynamics of ground water in the moss community station was closely related to the course of snow cover ablation in the upper part of Fugleberget drainage area. The ground water table, being in fact the water coming from the surface run off, during the snow cover ablation stayed at the depth of only 1 to 5 cm under the surface. In the period of intensive snow melting the water table stayed up to 2 cm over the surface of the area. In spring in this type of soil the level and dynamics of water table was not related to the permafrost level. Therefore, the share of permafrost waters in general water balance is small and water relations in soil profile are shaped by runing waters.

Humidity of the surface layer of peat soil during the whole period of spring snow cover ablation was formed by flowing ablation waters and amounted to 100% of maximal water capacity.

Refreezing rate of structural soil during the spring snow cover ablation was diversified. Some inhibitions and accelerations of this rhytm due to the air temperature changes were noted. In spring, in structural soil, quick and almost constant heat accumulation in soil profile took place. In June the permafrost was lowered to the depth of 71 cm and at the beginning of polar summer it stayed at the depth of 78 cm under the ground surface. The average twenty-four hours rate of soil refreezing was 2,1 cm; in the periods of retardation, i.e. at low air temperatures this rate diminished to 0,5 cm per twenty four hours; during the acceleration periods the rate increased to 3,5 cm per twenty four hours. In structural soil permafrost is an impermeable layer for ground waters. The level of the ground water table coming from the melting permafrost and from the infiltration of melted snow cover is related to the permafrost level. Ground water table quickly responded to the fluctuations of temperature in soil. Temperature increase caused a water drainage from the ground — the thickness of water layer in this period did not exceed 3 cm. When the temperature in soil decreased the water accumulated above the permafrost and then the thickness of water layer reached up to 13 cm. In the profile studied ground waters were not supplied by the running surface waters. The course of humidity changes of the surface layer of structural soil was closely related to the temperature and water regime of the soil (Fig. 5). At the beginning of studies the snow-melting water covering the ground surface caused that the soil

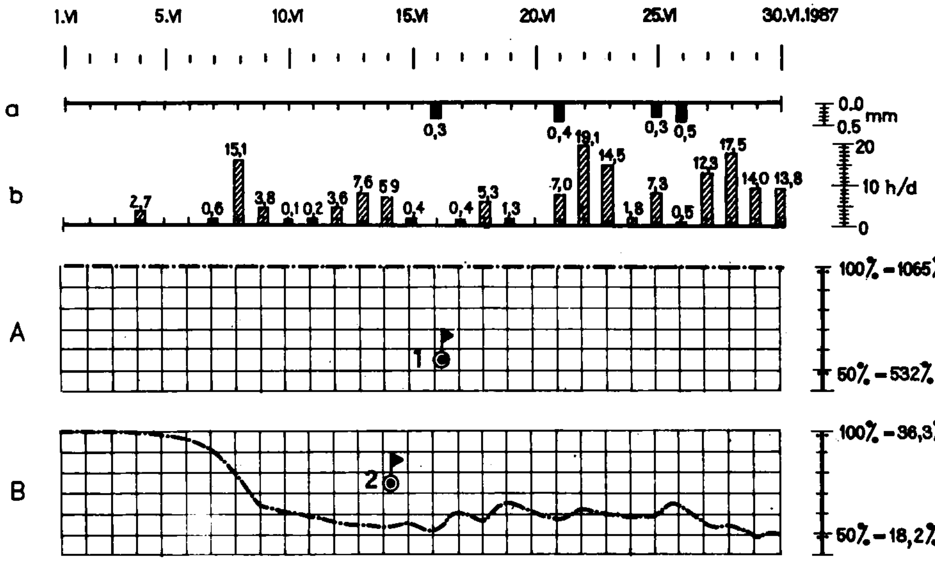

Fig. 5 Soil humidity (A — moss community, B — structural soil) in the surface layer at the depth of 5 cm in the period of snow cover ablation and precipitation (a) as well as isolation (b); Humidity is given in percent of full water capacity and in respective weight percentage.

was fully soaked with water. The run off from the ground surface caused a decrease of soil humidity, down to 90% of maximal water capacity. Evaporation from the soil surface and infiltration of water into the profile caused further decrease of soil humidity; in the first decade of June in the upper layer it reached a value of field water capacity (60% of maximal water capacity). Since then, the soil humidity balance was determined by precipitation and vapour condensation from one side and by evaporation from another. The lack of important precipitations caused that the soil humidity maintained within the range of 50—60% of maximal water capacity.

The thickness of wet layer related to the whole active layer was quite different in two soils studied; it depended on the quality and amount of water supply in the permafrost type of water relations. The wet layer in both soils amounted to from 8 to 100% of the whole active layer. However, in the case of peat soil situated on a way of snow cover ablation waters coming from the slope (erosion — lessivage type of water balance), it amounted to from 81 to 100% (Fig. 3). During the maximal intensity of run off of spring melting waters the soil of such water regime was under water for several days (in moss community the water table persisted above the ground surface for 6 days). In the case of structural soil with evaporation — hydromorphic type of water balance the wet layer amounted to from 8 to 50% of the whole active layer (Fig. 4). In the period of initial soil refreezing, the supply was determined by water coming from the snow cover ablation (infiltration) — then the wet layer in structural soil reached 50% of the whole active layer. When the supply of structural soil came from permafrost waters this quality was lower and ranged from 8 to 35%.

Conclusions

In similar meteorological conditions (i.a. lack of snow cover in spring) the thermic and humidity diversity of soils and the different rate of their refreezing during the polar spring are related to:

1. Different grain-size distribution, physical properties and morphology of soils resulting from their different origin. Diverse soils differently „manage" the heat received by their surface and thus are thermically different. Their refreezing rate is also different.

2. Different type of water balance. During the spring snow cover ablation, in permafrost type of water relations, local water regimes occur that are similar to the evaporation — hydromorphic type (structural soil) or to the erosion — lessivage type (peat soil). During polar spring such local water regimes decisively influence the surface layer humidity influencing also the soil thermics as well.

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Streszczenie

Wiosną 1987 w zlewni Fugleberget w rejonie fiordu Hornsund na Spitsbergenie (Rys. 1) prowadzono badania nad przepływem energii i obiegiem materii w układach glebowych tundry spitsbergeńskiej. Równolegle prowadzono obserwacje temperatury, wilgotności, dynamiki zwierciadła wody gruntowej i przebiegu odmarzania dwóch typów gleb. Badania objęły okres wiosennej ablacji pokrywy śnieżnej (czerwiec 1987). Do analizy wybrano gleby nizin nadmorskich: glebę strukturalną ze skrajnie ubogą roślinnością tundrową i mszarnik (zbiorowisko *Calliergon stramineum* — *Sanonia uncinata*) na powierzchniach prawie całkowicie pozbawionych pokrywy śnieżnej. Stwierdzono znaczne zróżnicowanie temperatury w profilach pionowych obu gleb, mimo zbliżonych warunków termicznych w obu siedliskach (Rys. 2). Mszarnik był glebą znacznie „chłodniejszą" od gleby strukturalnej (średnia miesięczna temperatura warstwy 0—15 cm wynosiła odpowiednio 3,0 i 4,7°C). Zróżnicowany był również przebieg i głębokość wiosennego odmarzania obu typów gleb oraz dynamika zwierciadła wody gruntowej (Rys. 3 i 4). Izolacyjny wpływ roślinności i wód pochodzących ze spływu topniejącego śniegu spowodował opóźnienie rozpoczęcia i zmniejszenie tempa odmarzania mszarnika w porównaniu z glebą strukturalną. Średnie miesięczne tempo odmarzania gleb wynosiło: mszarnika — 1,2 cm na dobę, gleby strukturalnej — 2,1 cm na dobę; 30 czerwca strop zmarzliny znajdował się na głębokości 41 cm (mszarnik) i 78 cm (gleba strukturalna). Gleby różniły się także wilgotnością warstwy przypowierzchniowej (Rys. 5). Średnia miesięczna wilgotność gleby w warstwie przypowierzchniowej (5 cm ppt) wyniosła: w mszarniku 100%, a w glebie strukturalnej 66% maksymalnej pojemności wodnej (odpowiednio: 1065 i 24% wilgotności wagowej). Powyższe różnice wypływają z odmiennej genezy gleb, a co za tym idzie — składu granulometrycznego, właściwości fizycznych i morfologii obu typów gleb oraz rodzaju gospodarki wodnej, jaki cechuje badane obiekty.