Studying the Dependence of Layer-by-Layer Snow Density Variability Dynamics upon Intra-Snow Crystal Formation in Snow Cover Evolution

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Abstract: Presentation of the results of testing a snow sampler with a controllable pitch designed for layer-by-layer snow pit studying. Demonstration of the snow cover structural features and diagenetic transformations typical for the winter of 2009-2010. Identification of the snow mass snow pack growth related to intra-snow crystal formation. Discussion of the mechanism of the process and its role in snow cover evolution. The results could be used for interpretation of the snow cover remote sensing data within zonal natural complexes–middle–taiga bilberry spruce forests.

Keywords: Snow cover evolution, subliminal metamorphism, snow layer texture, snow density, depth hoarfrost, Forel's lines, diagenesis of the snow mass, middle taiga, remote sensing.

1. INTRODUCTION

Nowadays snow cover remote sensing methods are being increasingly widely used [16, 24], and involving snow–measuring observations at network weather stations they provide the basic information on snow cover extent and its water equivalent [11]. Water equivalent can be measured remotely by the established empirical relationships between the amount of snow cover water equivalent and changes in the snow cover microwave radiation capacity [1, 6, 10, 27]. However, brightness temperature of the snow cover depends not only from its water equivalent [7] but also from the structure of the snow mass (snow depth and density; size of snow grains; wind compactions of snow (snow slabs); ice and radiation crusts; layer loosening and depth hoarfrost) [2, 3, 4, 32]. It is noted that thermal radio radiation of snow is strongly influenced by the depth and vertical in homogeneity of the snow cover [9, 13].

Meanwhile, because of highly uneven distribution of weather stations across most of the Earth's surface, there is no opportunity to characterize the snow cover structure and depth, as well as its water equivalent in advance. Therefore, tackling the inverse task–restoration of water equivalent and the structure of snow cover through modeling the snow mass stratigraphy by its radiophysical parameters–could reduce technical difficulties when evaluating the snow cover in remote and sparsely populated areas. The issue is complicated by insufficient regional knowledge of snow cover diagenetic transformations under various conditions of the snow accumulation, as it makes it difficult to interpret snow surface remote sensing results within some specific area–it is necessary to have data on the snow mass stratigraphy and textural irregularities [5, 29]. The lack of data in regional models of snow cover evolution necessitates using the average snow characteristics across the entire snow depth that results in divergence of experimental and theoretical results [11].

The article describes stratigraphic snow pits made in the snow mass within agro–landscapes (seeded grasslands with permanent grasses) established instead of original (primary) landscapes–middle–taiga bilberry spruce forests. These snow pits reflect the structure of the snow cover typical for the winter of 2009-2010, that is aligned with the regional conditions of snow cover formation and diagenetic transformations in the process of snow mass subliminal metamorphism in the period of snow cover growth.

2. STATEMENT OF PROBLEM

Subliminal metamorphism activity is considered to be displayed [17] through the structural transformations of snow layers (changes in granular texture of snow; formation of compacted/loosened zones in the snow mass). So, it can be assumed that layer–by–layer diagenetic transformations of snow layers could be reflected in vertical variation of the snow mass density. However, studying the dynamics of layer–by–layer snow density variability is not widely spread so far, probably because using cylindrical samplers for testing snow layers, as recommended in some papers [18, 25], leads to non–completely correct results, as a round profile of a horizontal core sample doesn't match the snow layer geometry under growing snow cover that makes the data less informative and comparable.

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E-ISSN: 2311-8741/15

Therefore, the work objective is to demonstrate the results of testing a snow sampler with a controllable sampling pitch when studying layer–by–layer diagenetic changes of the snow cover in the period of its growth.

3. CHARACTERISTIC OF SNOW ACCUMULATION

3.1. The Area of Interest

The area of interest is located in the watershed area in the pre–valley of the Sysola River and belongs to the Mezen–Vychegda Plain with steeply sloping and rolling topography. The primary vegetation is represented by middle–taiga green–moss bilberry spruce forests replaced in swampy interfluves by haircap–moss spruce forests with birches and pines. The landscape structure of watershed areas is mainly represented by combination of the following natural landmarks (natural boundaries): (a) flat poorly drained areas with marshy spruce forests on podzolic boggy soils; and (b) well–drained areas with green–moss bilberry spruce forests on podzolic soils. Instead of cut off primary forests there appear secondary mixed small–leaved forests with patches of tilled lands among them often covered by grassland vegetation.

3.2. Climatic Conditions of the Winter of 2009-2010

The late autumn was 1-3 degrees warmer as compared to the relevant multi–year data. In November the average temperature was –4…–9°C, and in 1st and 3rd five–day periods the temperature raised to +3…+9°C. The change from daily mean above–zero temperatures to stable daily mean subzero temperatures (the beginning of winter) took place on 15–17 November. The winter was hard. In the first decade of January the temperature fell to –33…–38°C. The second period of anomalously cold weather was recorded in 4th and 5th five–day periods of February when the temperature fell to –42…–46°C. The change to daily mean above–zero temperatures (the end of winter) took place on 28-29 March that was almost two weeks earlier than usual. Taking into account the winter's late onset and early end, the winter period lasted for 151-155 days that was four weeks shorter as compared to the relevant long–term observations.

3.3. Peculiarities of Snow Accumulation

The snow cover formed unevenly. The first snow fell in late October, and by the November thaw (the first decade) its depth was 18-23cm in the key area. By the end of the thaw the snow height had decreased by 4-6cm, and an ice crust (up to 3-5mm) formed on its surface which remained in the snow mass till the end of winter. The last five–day period of December and 1st half of January were characterized by heavy snowfalls– the snow depth increased by 10-20cm per decade. Then, till 24 February there were no many snowfalls, and the average weekly snow depth growth in the key area was 3-5cm. This period is characterized by the predominance of anti–cyclonic pattern of weather with rare snowfalls and low air temperatures. In the snowless period of January 13-25, a 2-3mm radiation crust of transparent ice formed on the snow surface which remained in the snow mass till the snow started melting. Since February 16, weekly snow cover observations were complemented by daily measuring the depth of the snow cover. Figure **1** demonstrates the dynamics of snow depth changes in the periods of its increase and loss. It is noted that the regime of snow accumulation has significantly changed by the end of winter as a result of heavy snowfalls at the end of February and in early March. In this period the snow depth growth amounted to 10-12cm. However, the maximum increase of the snow depth was recorded on 9-27 March. By the end of this period the snow mass increased to 81cm in depth. On 27 March warm weather set in, and the snow began to melt intensively, and by April 5 the snow depth had decreased to 42cm. The key area became completely free of snow on April 22, 2010.

Figure 1: The dynamics of snow cover growth and loss in the winter of 2009-2010.

4. METHODOLOGY

4.1. Observations of the Snow Cover Stratigraphic Variations

Observations of the snow cover stratigraphic variations were conducted from 25 January till 5 April 2010 in the field laid down with perennial grasses and located in the green zone 4 km. west of the City of Syktyvkar. They included weekly and since 16 February daily snow surveys with measuring the depth of the snow and describing snow pits through layer–by–layer snow sampling to define the snow density. Snow pits were performed in an upland zone with a one–week interval in the line across the direction of predominant snow–wind streams that prevented blowing of the snow to new snow–measuring sites. In the snow pits the snow mass stratigraphy and characteristics of the textural changes taking place during its growth were studied; the zones of subliminal congelation of snow grains (subliminal cementation, according to Shumsky, [23]) were visually determined. The latter was taken into consideration when estimating the level of snow moisture content in the contact zone between snow and soil. It should be noted that when estimating the snow depth with a snow measuring stick at the snow–measuring sites a subliminal cementation layer was not felt–it occurred only at the bottom of the snow mass and was recorded when selecting samples in the form of a layer of adfreezed depth hoarfrost crystals.

Along with describing the snow mass stratigraphy, schematic sketches of the front walls of the snow pits and microphotography of snow grains and subliminal clusters of depth hoarfrost ice crystals were also performed. Photo shooting was performed in the sun-light with Canon–A640 digital camera with 16X zoom, with the samples of snow taken from genetic horizons. For layer–by–layer snow sampling a special snow sampler with a controllable pitch was designed [21].

4.2. Description of a Layer–by–Layer Snow Sampling Device

Figure **2** demonstrates a layer–by–layer snow sampling device with a controllable snow mass sampling pitch. It consists of a right–angle prism which walls (1) are made of the material chemically inert to atmospheric components deposited in the snow mass. The walls of the prism (1) are fastened to stiffening ribs (2) which also act as guides when dipping the prism vertically in the snow mass. Snow is sampled with a removable front wall of the prism which acts as a cutter (3). The device has a cantilever platform (4) set on the front stiffening ribs, which allows to control a sampling pitch. The cantilever platform is provided with clamps (5) to fix it at the required depth.

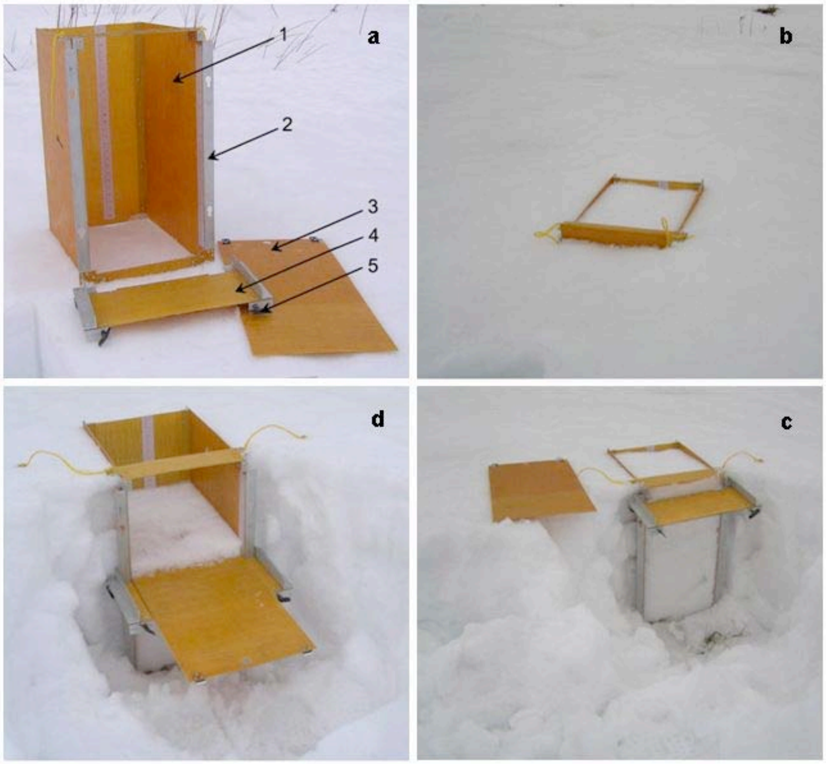

Figure 2: A layer-by-layer snow sampling device and the example of its application (the explanatory notes are given in the text).

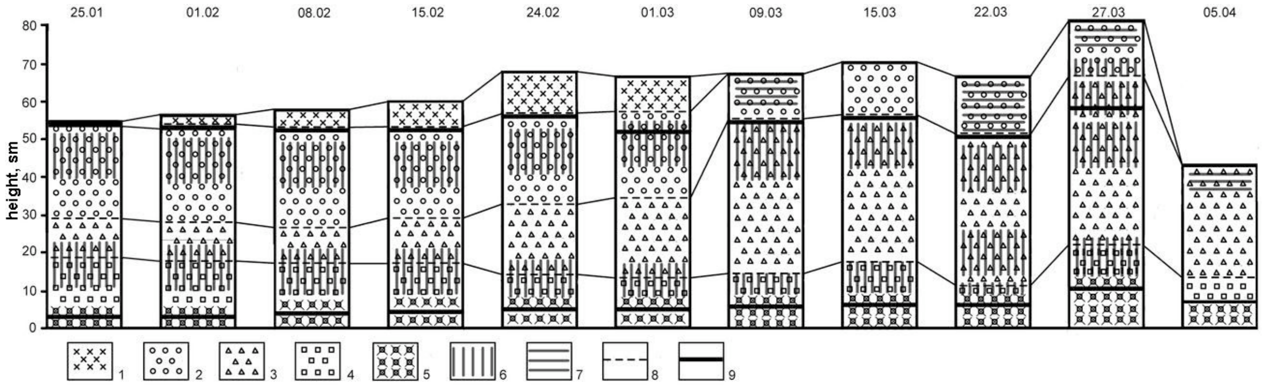

Figure 3: The dynamics of layer-by-layer snow cover variability, *Notation conventions: I – horizons (layers): 1 – new snow, 2 – fine-grained, 3 – medium-grained, and 4 – large-grained, 5 – depth hoarfrost crystals: II – density, 6 – loose zone, and 7 – compactions, Supporting notations: 7 – boundaries of horizons (layers), 9 – ice crusts.*

This device makes it technically possible to perform layer–by–layer snow sampling and change sampling pitches considering the structure of the snow mass and visually differentiating snow layers (in density and granular texture of snow; the presence of temperature (radiation) crusts; and wind compactions of snow (snow slabs)).

4.3. Layer–by–Layer Sampling of Snow for Density Measurement

To take samples a snow sampler should be dipped vertically in the snow mass (Figure **2, 6**). Then it is necessary to dig a snow pit so that the front wall of the prism (3) remains free of snow (Figure **2в**). The front wall of the prism (3) should be further removed, and the cantilever platform (4) should be fastened to the front stiffening ribs (2). Then it is necessary to select the required thickness of snow bricks (within the framework of these activities the interval of sampling remained unchanged – 2cm) and fix the platform at the required level with the clamps (5) (Figure **2, г**). To get a snow brick, the front plate of the prism (3) should be placed on the cantilever platform (4) and pushed steadily in the right–angle snow core bounded by the three walls of the prism. The snow brick produced (2 × 27 × 36cm) was placed in a plastic bag with a plastic scoop. The density of snow was measured on the days of taking samples. The samples were weighted within the accuracy of 0.01g., with the mean value recorded in the log.

5. RESULTS

5.1. Snow Mass Stratigraphy and its Variations in Snow Cover Evolution

By the start of observations (25.01.2010), the snow depth at the snow–measuring site was 54cm (Figure **3**). It should be noted that the beginning of research concurred with yielding of the frost (the air temperature raised from –30° to –17°C). However, the anti–cyclonic pattern of weather set in the first decade of January, lasted to February 8. In this snowless period a thin (1.5÷3mm) radiation crust of transparent ice occurred on the snow surface which then shifted deep down the snow mass as a result of the subsequent snowfalls. Radiation and thaw ice crusts stayed in the snow mass till the beginning of snow melting. In the process of snow depth growth observations the location of those ice crusts in the snow mass relative to each other turned out to be an informative marker to estimate the snow mass growth as a result of intra–snow crystal formation becoming more active during the passage of cyclones.

Along with the ice crusts described above, three fine–grained, medium–grained and large–grained snow horizons were identified in the snow mass. The upper layer up to 25cm thick was formed of loose white, opaque free–flowing fine–grained snow (Figure **4, а**) which was gradually changed by a medium–grained snow layer (10÷15cm) at the bottom (Figure **4, б**). At a depth of 35-54cm ice crystals became even more in their relative size–a depth hoarfrost layer was formed. Depth hoarfrost crystals were transparent and reached 3-5mm in size. They had razor edges and well–perceptible Forel's lines (Figure **4, в**).

It was found that depth hoarfrost crystals in the zone of contact between snow and soil were more moistened as compared to the snow grains in the upper layers. Thus, the snow brick produced at the bottom of the snow core sample (the snow pit made on 09.03) congealed in the open air and shaped into a plate (Figure **5, а**) of large depth hoarfrost crystals (Figure **5, б**). It was a rapid congelation–crystals froze together yet while on the cutter plate and took the form of a plate holding its shape even when being placed in the

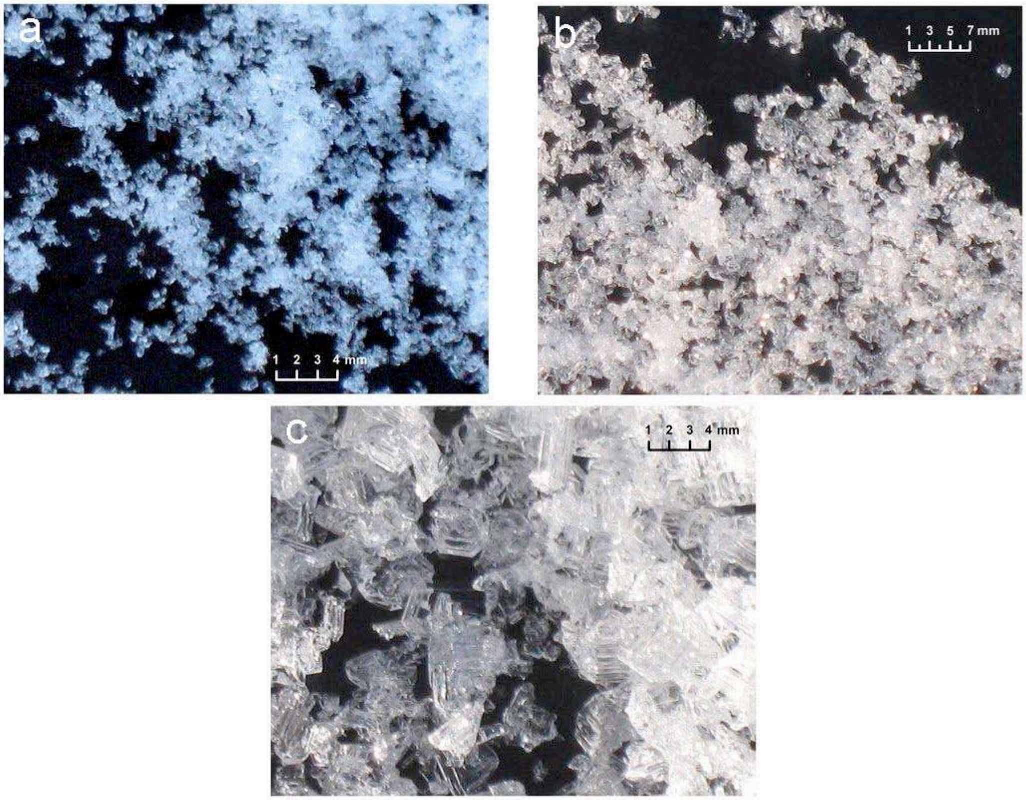

Figure 4: Microphotographs of snow grains and ice crystals, (**a**) fine-grained snow, (**б**) medium-grained snow, (**в**) large-grained snow represented by large (up to 3-5mm) depth hoarfrost ice crystals with Forel's lines.

packet to thaw. However, some changes in the layer's limits should be noted. While the lower limit was relatively even (Figure **5, а**), the upper limit had some asperities in contrast to the previous sample. The asperities were formed later–during a short period of describing the previous sample–and they were formed by separate depth hoarfrost grown crystals (Figure **5, б, в**). It is assumed that the crystals' rapid growth was conditioned by the excess interstitial water contained in the snow horizon with depth hoarfrost. It was also noticed that the depth of the snow horizon characterized by subliminal adfreezing (cementation) of layer–by–layer snow samples depended upon the height of the snow cover. The snow cover height being 40-50cm, the depth of the snow horizon was 4-5cm, and the snow cover height being 70-80cm, the depth of the snow cover increased to 10cm (Figure **3**).

It should also be noted that in the period of snow mass growth layers formed by snowfalls were not visually delimitable on the snow pits' front walls. Only at the top of snow profiles there were short–term layers formed by stratigraphically significant snowfalls, which then consolidated in one mass with an underlying fine–crystalline snow layer. In the middle and lower parts of the snow pit formed by medium–grained and large–grained snow there were no visible snowfall–related textural changes defined.

Figure 5: Subliminal adfreezing of depth hoarfrost crystals into a plate (**а**), formation of asperities on the plate in the process of depth hoarfrost crystal growth (**б, в**).

On the whole, this three–layer snow mass structure with two ice crusts remained unchanged till the end of the snow accumulation period (till 27.03). When intensive snow melting started, a three–layer snow mass structure transformed into a two–layer structure consisting of medium–grained and large–grained snow horizons, and the ice crusts melted.

5.2. When Studying the Dynamics of Layer–by–Layer Snow Density Variations

In the period of snow cover growth, the snowpack growth between two ice crusts was identified. It became noticeable when cyclones and heavy snowfalls arrived and was recorded by changes in the distance between radiation and thaw crusts in the middle of the snow mass. (Figure **6**, the snow pits made from 24.02 till 27.03). However, the average layer–by–layer snow density within this interval remained relatively constant.

When observing the dynamics of layer–by–layer snow density variations, there was revealed a trend of snow density growth with increasing depth (Figure **6**), which remained unchanged until the snow started melting. By the end of the snow accumulation period and under increasing cyclonic activity layer–by–layer snow mass density variations had become more noticeable. So, if in February snow density variations in the middle of the snow mass were within the interval of 0.20-0.25g/cm^3 (the snow pits made from 08.02 till 24.02), then in March the range of variations increased to 0.20-0.35g/cm^3 (the snow pits made from 01.03 till 27.03). In the last decade of March relatively warm and moist air came accompanied with increased wind that led to compaction of both the snow surface from 0.09 to 0.22g/cm^3 (the snow pits made from 09.03 till 22.03) and the upper part of the snow mass to a depth of 16-20cm where it increased to 0.25g/cm^3. As early as in a week the density of snow layers changed. In the snow pit made on 27.03, under the upper compacted layer (0.20g/cm^3) of fine–crystalline dry loose (free–flowing) snow there occurred a layer of medium–crystalline dry loose (free–flowing) snow with the depth of up to 10cm and the density of 0.13-0.15g/cm^3 lying on a compacted snow layer which earlier had loose structure (Figure **6**). The growth of the snow depth led to the reduction of snow density variations in the lower part of the snow mass.

On the whole, it should be noted that in the period of snow cover growth the density of new snow in the snow pits made up 0.07-0.11g/cm^3, while the snow density in deep layers was 0.29-0.40g/cm^3. At the same time the average snow density between the snow pits did not vary significantly–from 0.22 to 0.24g/cm^3. In other words, vertical (layer–by–layer) snow density variations in the snow pit have greater contrast and they are more dynamic as compared to the average snow density indices between the snow pits.

6. DISCUSSION OF RESULTS

6.1. Intra–Snow Crystal Formation in the Surface Snow

The snow mass increases in volume without significantly changed density of upper snow layers due to the increased interstitial space and the level of a liquid film on the surface of snow grains which is formed from condensed water vapour coming in the upper snow layers from the surface air. As a result of this mechanism, intra–snow crystal formation takes place which leads to the snow mass growth in the upper part of the snow mass.

This concept has the following basis. Snow cover is a porous two–component medium (air–water) [18, 24], where pore walls are formed by ice crystals of snow. As noted, because of constantly ongoing sublimation processes, snow pores, in contrast to other porous

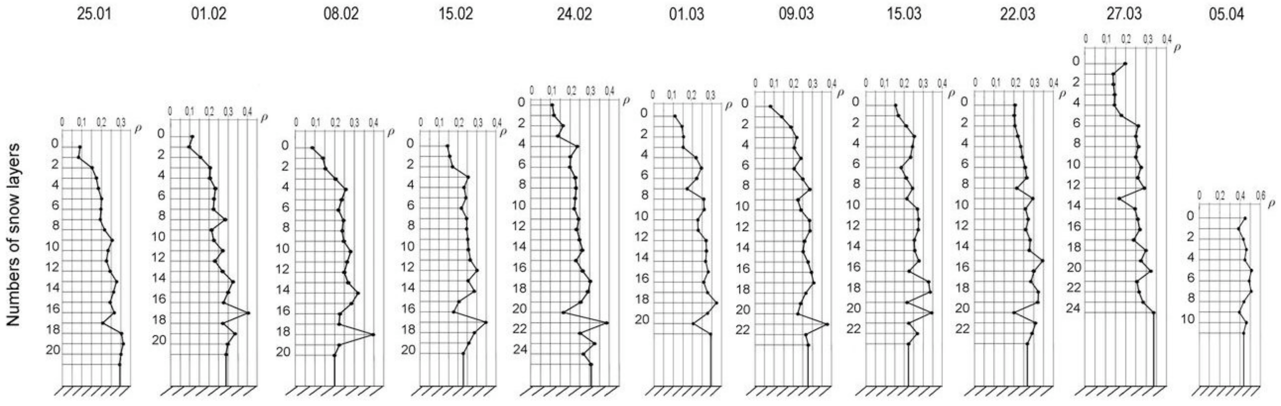

Figure 6: The dynamics of layer-by-layer snow mass density variability in the period of snow cover growth.

substances, are always saturated with water vapour [20]. Its condensation leads to formation of a sub–cooled liquid film covering the surface of large pores and forming capillary liquid in small pores. Thus, under increasing relative air humidity, snow is capable to take up water from the air and retain it in the snow mass due to capillary condensation. The latter is assumed to grow when a cyclone arrives. Cyclone air masses are characterized by comparatively high relative humidity and temperature as compared to the surface air above the snow surface[1], and vertical air mixing is a result of the turbulence conditioned by the snow inversion [22] which leads to heat and water exchange between the surface air and the snow surface. As a result, in the surface air there occur turbulent vortexes, and air mixing takes place which intensity depends upon the temperature lapse rate [28]. It is known that turbulence conditioned by temperature differences between the snow cover and the surface air could be accompanied with an organized convection when there occur up casts and downcasts. The latter are less intensive but cover large areas [14]. This vertical turbulent exchange results in a concentration gradient of atmospheric moisture. As saturated vapour pressure above the snowy surface is lower than that in the surface air, water vapour starts moving to the snow cover that leads to the growth of the relative humidity of the upper part of the snow mass.

In the context of contrast temperatures, as a result of arriving cyclones, a downcast with a relatively high water vapour content acts as a peculiar pump injecting excess water from the air into the snow cover, and an intra–snow crystal formation mechanism is started up. It works in the following manner. Incoming water vapour condensates on the pores' walls and, therefore, increases the thickness of a liquid film. The existence of water liquid phase at temperatures below freezing is not an extraordinary phenomenon. Possible formation of a sub–cooled liquid film on the surface of ice crystals is demonstrated theoretically and experimentally by B.P. Veinberg [31] and V.I. Kvilividze *et al*. [19], respectively. It is assumed that gradient (thermophoretic) motion of a liquid film over the surface of some ice crystal edges in the context of interstitial space saturated with water vapour could lead to their growth and increased interstitial space that will facilitate incoming of extra water vapour from the surface layer in this part of the snow mass. In the long run, this initiates intra–snow crystal formation which results in the increased volume of the snowpack in the snow cover.

6.2. Undulating Changes in the Snow Density

It is assumed that intra–snow crystal formation in the upper part of the snow mass takes place due to incoming extra water vapour from the surface layer what is confirmed by undulating changes in the snow density with a relatively stable trend of its growth down the snow profile (see Figure **3**). This assumption is based on the research of water vapour diffusion in the snow cover [26]. In this work a well correlation between the snow temperature and density undulating distribution and the layered distribution of snow mass humidity is demonstrated. Moreover, there are known experimental studies [28] confirming the dependence of vapour diffusion in the upper part of the snow mass upon the temperature gradient, where the gradient contrast was more dependant upon the wind pattern rather than upon the snow surface temperature dynamics.

6.3. Intra–Snow Crystal Formation in the Contact Zone between Snow and Soil

The mechanism of snowpack growth at the bottom of the snow mass is different and conditioned by an isothermal layer located in this zone. It is know that in the contact zone between snow and soil the course of isotherms significantly changes [12]. It is conditioned by lowering soil temperature as a result of lower snow layers cooling caused by water vapour evaporation from the soil. The density of soil and its volumetric heat capacity are 5 times as high as those values of snow, and soil in the contact zone with snow is cooled far less as compared to the soil in the surface snow. Therefore, in the process of soil water evaporation in the contact zone between snow and soil the temperature of snow lowers more rapidly than the temperature of soil. As a result of this, there occurs an isothermal layer [12] where depth hoarfrost crystal moisture is higher than in overlying large–grained snow layers. So, in open snow pits isothermal layers are characterized by formation of subliminal cementation zones. Higher moisture content of ice crystals in this zone is likely conditioned by excess water vapour occurred in the process of its nonlinear mass transfer from the soil within the isothermal layer, which partly condensates on depth hoarfrost crystals in the form of a sub–cooled liquid film which, in its turn, crystallizes on their edges and increases the size of crystals, and could lead to

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[1] Snow surface is characterized not only by high reflecting and emissive capacity, but also dehydrates the surface air [22].

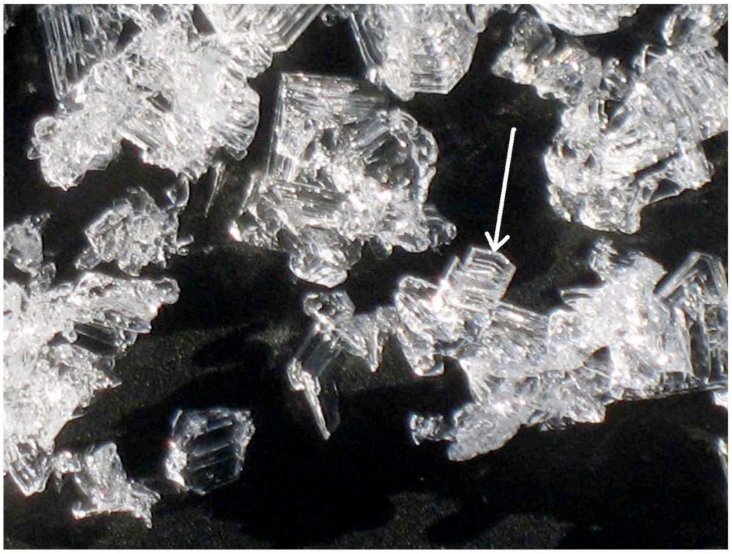

Figure 7: Depth hoarfrost crystal morphology (the example of a skeleton crystal form is shown by the arrow) within the isothermal layer zone.

formation of peculiar skeleton crystal forms (Figure **7**). It should be noted that such skeleton crystal forms morphologically differ from similar formations mentioned in this work [15]. These new skeleton crystal formations may participate in increasing the interstitial space volume in the depth hoarfrost zone due to incoming extra water vapour from the upper soil layers that eventually results in the increased snowpack volume at the bottom of the snow pit. Therefore, within the isothermal layer there is a peculiar crystal–forming medium which, in a certain manner, influences the dynamics of snow accumulation at the bottom of the snow mass.

6.4. Forel's Lines on the Surface of Depth Hoarfrost Ice Crystals

Indicate eutectic equilibrium zones between the solid states of ice and salt solutions in the form of interpenetrating areas, which constitute a two–phase system–cryohydrates. This assumption is based on the following. Hydroscopic water soluble particles are known to be able to act as condensation and sublimation nucleus in the atmosphere. Participating in aerosol formation, they largely define the precipitation composition. So it is assumed that in upper snow cover layers capillary liquid on the surface of ice crystals and the mineral composition of condensation nucleus are saturated solutions and are at equilibrium. According to Raoult's law, a solute's congelation point is known to lower proportionally to its molar concentration–the congelation point of such solutions is below 0°C. When the system temperature passes over 0°C, its components start interacting in a low–temperature mode and remain liquid (sub–cooled). Thermodynamic constants of reacting compounds remain effective, as thermodynamics does not consider rates of reactions. It only defines which form is more stable in these conditions.

As a result of temperature lowering, cryogenic dehydration of the solution takes place when a crystal phase–volumetric ice–and a liquid phase–concentrated aqueous solution of salt–are separated. Subsequent freezing of the solution leads to the formation of a solid eutectic mixture–cryohydrates which are a two–phase system consisting of a thin mixture of salt and ice crystals. Clear ice and a eutectic mixture are assumed to have different optical density, and lines become discernible in reflected light–Forel's lines. In other words, on the edges of depth hoarfrost crystals these lines indicate eutectic equilibrium zones between the solid states of ice and salt solutions in the form of interpenetrating areas, which constitute a two–phase system–cryohydrates. To accomplish nucleation and to reach the eutectic equilibrium, very slow cooling is required, and these particular conditions exist in the isothermal layer within which there are no significant temperature gradients. However, in the zone of depth hoarfrost formation the eutectic equilibration could be disturbed by excess water vapour coming from the soil. It condensates on the surface of ice crystals and, therefore, influences the kinetics of crystal growth which, in its turn, impacts the visibility and spacing of Forel's lines. Its persistence on various crystals (see Figure **6, c** and Figure **7**) may confirm a constant inflow of water vapour from the soil to lower snow layers and a constant process of intra–snow crystal formation. Therefore, intra–snow crystal formation at the bottom of the snow profile, which role in snow cover evolution is poorly studied and is not considered in snow cover modeling, can be an important source of snow formation [8].

7. CONCLUSION

Within zonal natural complexes represented by middle–taiga bilberry spruce forests a device for layer–by–layer snow mass sampling was experimentally tested. A technical result achieved by means of this device is an opportunity to perform layer–by–layer snow sampling with a controllable pitch depending upon the structure of the snow mass and the presence of visually differentiating snow layers (in density and granular texture of snow; the presence of temperature (radiation) crusts; and wind compactions of snow (snow slabs).

In the snow mass there was revealed a trend of snow density growth with increasing depth, which remained unchanged until the snow started melting. At the same time in the process of snow depth growth layer–by–layer snow density variations in the snow pit (space factor) started to have greater contrast, while the average snow density between the snow pits (time factor) did not vary significantly in the period of snow cover growth. In this context, it would be reasonable to study changes in these values during the loss of snow cover.

Visual delimitation of the snow cover layers related to stratigraphically significant snowfalls was found to be impossible. Only at the top of the snow profile there were short–term layers with textural variability, which then consolidated in one mass with an underlying fine–crystalline snow layer. In the middle and lower parts of the snow mass there were no visible snowfall traces at all. On the whole, the snow cover had a three–layer structure formed by fine–crystalline, medium–crystalline and large–crystalline snow layers. In the winter of 2009-2010 such snow mass stratigraphy remained unchanged during the entire period of snow cover growth. When intensive snow melting started, a three–layer snow mass structure rapidly transformed into a two–layer structure formed by medium–grained and large–grained snow.

It is found that diagenetic snow mass transformations are accompanied with the growth of snowpack due to intra–snow crystal formation. Secondary formations–radiation and thaw ice crusts–are not eliminated in the snow mass and remained there till the snow starts melting. It was demonstrated that secondary formations–radiation and thaw ice crusts–in the snow mass could indicate the snowpack growth in the process of intra–snow formation.

Forel's lines on the surface of depth hoarfrost ice crystals are assumed to indicate eutectic equilibrium zones between the solid states of ice and salt solutions constituting a two–phase system–cryohydrates. It is also assumed that clear ice and a eutectic mixture formed by interpenetrating areas of the crystalline solute and ice have different optical density. Therefore, on the edges of depth hoarfrost crystals this zone becomes visible in the form of alternate lines–Forel's lines indicating the zones of ice crystal growth.

When evaluating the current state of the snow cover stratigraphy, meteorological parameters of arriving air masses (temperature and relative air humidity) should be considered as they differ significantly in “cold” and “warm” winters [30]. It is assumed that increasing cyclonic activity will result in the increased share of intra–snow crystal formation and the relevant snowpack growth in snow accumulation balance that could, in a certain manner, influence snow radio brightness parameters in the process of its surface remote sensing within zonal natural complexes – middle–taiga bilberry spruce forests.

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Received on 14-08-2015 Accepted on 10-09-2015 Published on 31-12-2015

DOI: http://dx.doi.org/10.12974/2311-8741.2015.03.01.2