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Subsurface fluxes of mass and energy at the accumulation zone of Lomonosovfonna ice cap, Svalbard

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Abstract

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Glaciers cover ca 10% of the Earth's land and are found in the high altitudes and latitudes. They are important components of environmental systems due to the multiple feedbacks linking them with the atmosphere, hydrosphere and periglacial landscapes. The cold sloping surfaces of glaciers change the patterns of atmospheric circulation at different scales and at the same time glaciers are largely controlled by climate. They are commonly used as climatic archives for reconstruction of the past environmental changes based on evidences from the areas affected by glaciation at the moment and in the past. Glaciers are the largest fresh-water reservoirs on our planet and runoff thereof significantly affects the global sea level and life in glaciated catchments. However, melt- and rain-induced runoff from glaciers greatly depends on the subsurface conditions which thus need to be taken into account, particularly in a changing climate.

This thesis focuses on the processes of subsurface mass and energy exchange in the accumulation zones of glaciers, which are largely driven by the climate at the surface. Results are largely based on empirical data from Lomonosovfonna ice cap, Svalbard, collected during field campaigns in 2012-2017. Observations of subsurface density and stratigraphy using shallow cores, video records from boreholes and radar surveys returned detailed descriptions of the snow and firn layering. The subsurface temperature data collected using multiple thermistor strings provided insights into several subsurface processes. The temperature values measured during three summer seasons were used to constrain the suggested parameterization of deep preferential water flow through snow and firn. The part of data recorded during the cold seasons was employed for an inverse modelling exercise resulting in optimized values of effective thermal conductivity of the subsurface profile. These results are then used to compute the subsurface water content by comparing the simulated and measured rates of freezing front propagation after the melt season in 2014.

The field observations and quantitative estimates provide further empirical evidences of preferential water flow in snow/firn packs at glaciers. Results presented in the thesis call for implementation of description of the process in layered models simulating the subsurface fluxes of energy and mass at glaciers. This will result in a better understanding of glacier response to the past and future climatic changes and more accurate estimates of glacier runoff.

Keywords: glacier, ice sheet, sea level, runoff, ice, firn, snow, stratigraphy, density, core, radar, thermistor, temperature, preferential water flow, thermal conductivity, water content

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To Lena who inspires me by what and how she is and does

List of papers

This thesis is based on the following papers, which are referred to in the text by their Roman numerals.

I Marchenko, S., Pohjola, V., Pettersson, R., van Pelt, W., Vega, C., Machguth, H. and Isaksson, E. (2017). A plot-scale study of firn stratigraphy at Lomonosovfonna, Svalbard, using ice cores, borehole video and GPR surveys in 2012–14. Journal of Glaciology, 63(237), 67–78. doi: 10.1017/jog.2016.118

II Marchenko S., van Pelt W, Claremar B., Pohjola V., Pettersson R., Machguth H. and Reijmer C. (2017). Parameterizing deep water percolation improves subsurface temperature simulations by a multilayer firn model. Front. Earth Sci. 5:16. doi: 10.3389/feart.2017.00016

III Marchenko S., van Pelt W, Pohjola V., Pettersson R., Lötstedt, P. and Reijmer C. (2018). Thermal conductivity of firn at Lomonosovfonna, Svalbard, derived from subsurface temperature measurements. *Manuscript.*

IV Marchenko S., van Pelt W, Pohjola V., Pettersson R., Lötstedt, P. and Reijmer C. (2018). Water content of firn at Lomonosovfonna, Svalbard, derived from subsurface temperature measurements. *Manuscript.*

Front cover of the printed version of the thesis shows ablation zone of the Root glacier, Wrangell mountain in Alaska, the USA. The photograph is made on the 5^{th} of June 2010 and shows a supraglacial stream - a river carrying melt water downslope on top of ice.

Back cover of the printed version of the thesis shows an englacial water drainage channel - a tunnel formed by discharge of a large volume of melt water through ice. The photograph was made on the Root glacier on the 11^{th} of June 2010.

Contents

1. Background

1.1 Glaciers as components of environmental systems

Glacier is a perennial mass of ice, and possibly firn and snow, originating on the land surface from snow or other forms of solid precipitation and showing evidence of past or present flow (Cogley et al., 2011). It is important to state that in terms of surface mass fluxes glaciers can be divided in two domains. Accumulation zones are typically found at higher elevations where precipitation rates is higher and melt is less effective which altogether results in a positive net annual surface mass flux. Ablation zones lie at lower elevations and are consequently dominated by mass loss either by melt or by calving of icebergs. The resulting spatial imbalance of the mass fluxes is counteracted by viscous ice flow driven by the gradient of the overburden pressure and directed downslope from accumulation towards ablation zone.

Existence of glaciers requires low air temperature with a steady input of solid precipitation. Such conditions are characteristic of high altitudes and latitudes regions on the Earth. Currently the Earth's two ice sheets (Antarctic and Greenland) and *ca* 200000 glaciers cover respectively *ca* 13.5, 1.7 and 0.7 mln km^2 or *ca* 3% of the Earth's surface (Benn and Evans, 2010; Pfeffer et al., 2014). Glacier ice is the largest fresh water reservoir on the Earth and the hypothetical melt of the entire glacier mass would result in the increase of the global sea level by *ca* 64 m (Vaughan et al., 2013).

The significant scale of the Earth's glaciation and its tight links with the atmosphere, hydrosphere and periglacial environments at different scales explain the need for the humankind to study glaciers. The state of glaciers largely relies on the amount of precipitation and melt which are largely controlled by the processes in the lower atmosphere. In their own turn, processes at and below the surface of glaciers constrain the lower boundary conditions for the atmosphere. Large surfaces of glaciers are covered by snow with high albedo and therefore reflect much more solar radiation back towards atmosphere than water or most landscapes (for example vegetation). Furthermore, the ice surface temperature can not exceed $0°C$. These two phenomena along with the fact that often significant ice thickness rises the glacier surfaces at higher elevations encourage existence of relatively cold, vertically stable atmosphere over glaciers. That has such important implications as existence of large-scale air pressure maxima over ice sheets affecting the global patterns of atmospheric circulation (Parish and Bromwich, 2007; Steffen and Box, 2001) and katabatic winds observed over valley glaciers (Obleitner, 1994).

The primary mechanisms of mass wastage from glaciers are ice melt and calving of icebergs. Glacier ablation defines discharge of rivers in glaciated catchments affecting the valley ecosystems (e.g. Bliss et al., 2014; Milner et al., 2017) and is often an important source of water supply for agriculture (e.g. Juen et al., 2007) and hydro-power plants (e.g. Farinotti et al., 2016). Mass wastage from glaciers and ice sheets in one of the most significant contributors to the change of the global sea level (e.g. Vaughan et al., 2013) which has global implications and a high practical significance. Furthermore, the mass of ice sheets is significantly large to cause deformation of the Earth's crust and displacement of the underlying mantle altogether causing a change in the elevation of land surface relative to the sea level in response to climatic changes (Johansson et al., 2002).

Glaciers are widely used as climatic archives. External environmental perturbations find reflection in the physical and chemical properties of the accumulation layers which can thus be used as proxies to reconstruct the past climatic changes. For example, the isotopic composition of glacier ice samples from deep cores drilled at the Greenland (e.g. Johnsen et al., 2001) and Antarctic (e.g. Petit et al., 1999; EPICA community members, 2004) ice sheets allowed to reconstruct the climatic and atmospheric history at the two ice sheets during the last 110 and 740 thousand years respectively.

Glaciers exert a dramatic effect on the landscape through erosion, transport and subsequent deposition of the bedrock and sediments by the flowing ice and water masses. Thus advancing and retreating glaciers alter the periglacial landscapes. That allows to reconstruct the environmental conditions contemporary to the corresponding changes basing on the empirical evidences from the moraine complexes and sediments found both on land (e.g. Bennett and Glasser, 2009) and on the sea bottom (e.g. Bond et al., 1992; Dowdeswell et al., 1998).

There are also such practical aspects of glaciology as traveling in glaciated areas and the associated risks, glacier-related outburst floods (Björnsson, 2003; Gagliardini et al., 2011), ensuring the safety of navigation in seas affected by iceberg calving (Bigg and Wilton, 2014), mining in glaciated areas (Kronenberg, 2013) and pollution of glaciated environments (Colgan et al., 2016). Finally, but not least important is that glaciers provide a profound source of fascination and form the core of magnificently beautiful landscapes (e.g. Post and LaChapelle, 2000; Holmlund and Jansson, 2003). Thus it is important to understand the processes within glaciers, what governs their changes and what implications for the environment can be expected from these changes.

1.2 Mass and energy fluxes at glaciers

The present thesis is focused on the subsurface processes in the accumulation zones of glaciers which are in close relation with the mass and energy fluxes at the surface. The present chapter intends to provide the background information on the topic. First mass and energy fluxes at the surface are considered. That is followed by a review of principal subsurface processes along with an overview of published approaches for describing them.

1.2.1 Processes at the surface

According to the Glossary of mass balance and related terms (Cogley et al., 2011, see Figure 1.1) the main source of mass at a glacier's surface is solid atmospheric precipitation coming either as snow, rime (freezing of supercooled water) or deposition (phase shift of H_2O from vapor directly into solid). Another possible contributors are: wind drift, avalanching and freezing of water melted earlier or at a different location and then transported to the site in question. The main mechanism of mass wastage at the surface is ice melt followed either by infiltration of the liquid water downwards along the pores of underlying snow and firn or lateral runoff on top of a sloping ice surface. However, ice can be also sublimated by the solid to vapor phase shift (Lliboutry, 1954). Glaciers terminating in sea or lake waters or on extremely steep slopes also loose mass through calving of icebergs (Benn et al., 2007).

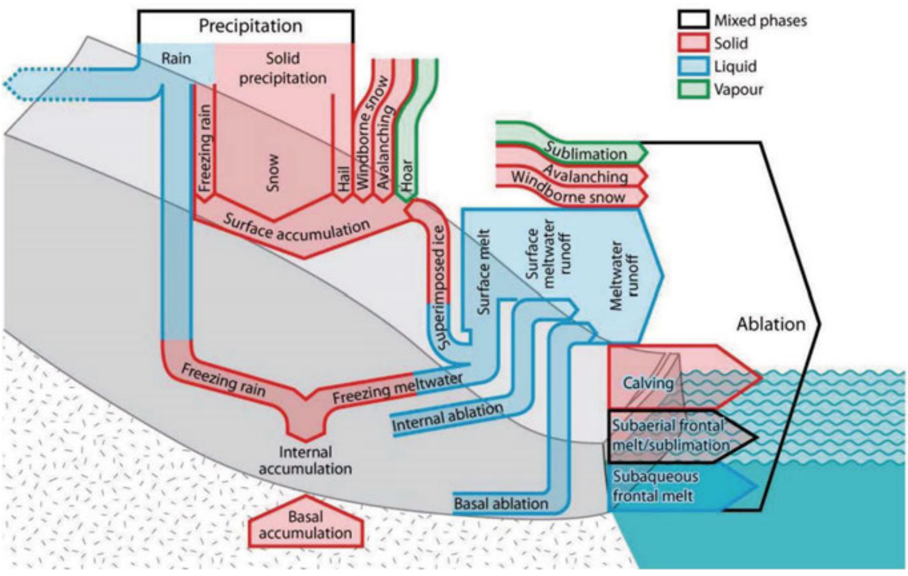

Figure 1.1. Components of the mass balance of a glacier. The arrows have arbitrary widths and do not indicate physical pathways of mass transfer. (Cogley et al., 2011)

The fluxes of energy at the surface can be described by the following equation (e.g. Oerlemans, 2010; Cuffey and Paterson, 2010):

$$SW\downarrow + SW\uparrow + LW\downarrow + LW\uparrow + H_s + H_l = G + M. \tag{1.1}$$

The down-welling solar short wave radiation ($SW\downarrow$) is the primary source of energy and is controlled many factors, among which are time of the year and day, local latitude, slope aspect, shading from the surrounding terrain and optical thickness of the atmosphere. The latter depends on the cloud conditions, air humidity and aerosol concentration.

The up-welling short wave radiation is ($SW\uparrow$) linked with $SW\downarrow$ through albedo (α): $SW\uparrow = \alpha \cdot SW\downarrow$. Fresh snow albedo is usually high (*ca* 0.9) suggesting that almost all short-wave radiation impinging on the glacier surface is reflected back to the atmosphere. Over time α decreases but usually remains above 0.5 for snow and above 0.3 for ice. Albedo is a crucial parameter for the surface energy balance as it has a pronounced effect on the short-wave radiation balance.

The downwards and upwards long-wave radiation fluxes ($LW\downarrow$ and $LW\uparrow$) are defined by the temperature of the lower troposphere and glacier surface respectively. The glacier surface temperature (T_s) can not exceed $0°C$, which according to the Stefan-Boltzmann law effectively limits the upwelling long-wave radiation at $LW\uparrow = \sigma \cdot T^4 \approx 316\ W\ m^{-2}$. Cloud cover increases $LW\downarrow$ and thus on a cloudy day the net long-wave radiation balance at a melting surface is usually positive.

The sensible (H_s) and latent (H_l) heat fluxes are commonly referred to as turbulent heat fluxes as both largely depend on the wind turbulence in the lowermost atmosphere layers. The fluxes are respectively driven by the vertical gradients of air temperature (H_s) and humidity (H_l).

The term G refers to the downwards energy flux in an infinitely thin layer just below the surface. It can be caused by penetration of solar radiation and by the thermal conduction. The latter is driven by the temperature gradient and depends on the thermal conductivity of the subsurface profile. If the sum of terms on the left hand side of Equation 1.1 is positive and $G = 0$ the energy M is used for melting the surface. The rate of surface lowering (c_{sfc}) due to melt can be expressed as: $c_{sfc} = \frac{M}{L \cdot \rho}$, where L is latent heat of ice melt and ρ is the subsurface density.

The rate of surface accumulation and melt at a point can be estimated using density measurements along with stake readings, probing, sonic ranger or other remote sensing techniques (Cogley et al., 2011). However, instrumental measurements at glaciers are logistically challenging and not always possible or accurate enough. As a result measurements done outside glaciers are often extrapolated to yield precipitation and melt rate estimates for glaciated catchments (e.g. Van Pelt and Kohler, 2015). Equation 1.1 is commonly used to estimate M from measurements and/or parameterizations of the associated terms (Hock, 2005). Under a lack of data to constrain a full energy balance calculation air temperature is commonly used to parameterize glacier melt rate due to its involvement in multiple feedbacks (Ohmura, 2001). The recent advances in process-understanding and efficiency of computer calculations made it possible to include detailed descriptions of energy and mass fluxes at glaciers in regional climate models (e.g. van de Berg et al., 2006; Fettweis, 2007).

1.2.2 Subsurface processes

The conditions observed below a glacier surface are the result of several mass and energy exchange processes with the largest contributions coming from the following phenomena: refreezing of melt and rain water (e.g. Reijmer et al., 2012); conductive heat flux (e.g. Sturm et al., 1997); gravitational settling (Lundin et al., 2017) and metamorphism (e.g. Lehning et al., 2002) of porous snow and firn. These processes are linked with each other and are to a large extent controlled by the surface conditions. The uncertainties associated with the latter are translated to the formulations of the surface mass and energy fluxes.

The fact that rain and melt water does not necessarily contribute to runoff from a glaciated catchment, but can be refrozen or retained within a glacier was acknowledged quite early (e.g. Hughes and Seligman, 1939). However, accounting for these processes in estimates of the climatic glacier mass balance remains a challenge. Insufficient understanding of subsurface processes and a lack of empirical data on the parameters controlling these process complicates accurate estimates of subsurface water refreezing and retention rates. Water infiltrating through cold snow and firn refreezes in pores releasing the latent heat of fusion L equal to 333500 $J\ kg^{-1}$ (Cuffey and Paterson, 2010), which makes it a very efficient way to rise the subsurface temperature and density. Furthermore, capillary forces can retain some water in the pores, which is likely to be refrozen later during the cold season. Surficial water can also freeze on top of cold impermeable ice surface and form superimposed ice. Below several approaches to quantification of water refreezing in glaciers are reviewed.

Reeh (1989) assumed that the fraction of melt water refreezing annually (P_{max}) equals to 0.6 of the annual precipitation. The value for P_{max} was chosen arbitrarily to match simulation results with other estimates or runoff from the Greenland ice sheet.

Oerlemans (1991) assumed that in Equation 1.1 melt rate M is equal to runoff and the energy G is spent on changes in temperature T_{2m} of the glacier layer equivalent to a 2 m thick ice slab both by conductive heat transfer and refreezing. When the sum $B = M + G$ is positive M is defined as an exponent function of T_{2m} expressed in [$^{\circ}C$]: $M = B \cdot e^{T_{2m}}$. Oerlemans (1991) does not explicitly describe the relation between G and T_{2m} but mention that coefficients involved in calculations are chosen to match the results with earlier findings.

The fact that refreezing of the surficial water at glaciers can be limited either by the refreezing capacity (RC) of the subsurface profile or by the pore space therein (P), a number of authors suggested to use one of the two (Braithwaite et al., 1994; Huybrechts and de Wolde, 1999) or both (Shumskii, 1955; Pfeffer et al., 1991; Janssens and Huybrechts, 2000) constrains to quantify the difference between melt and runoff rates.

The zero-runoff condition regarding refreezing capacity for a snow/firn/ice layer with temperature T, density ρ and the depth of zero annual temperature amplitudes z_0 can be formulated as:

$$\int_{z_s}^{z_0} CT\rho dz > WL \tag{1.2}$$

where z is depth, z_a is the surface, C is the heat capacity of ice, W is the summed mass of melt and rain water. To ensure that the subsurface pore space is not entirely filled by refrozen melt water over time a certain relation between the accumulation and ablation rates is necessary:

$$\frac{\rho_d - \rho_u}{\rho_d} \cdot S > W. \tag{1.3}$$

Here S is the mass of solid precipitation, ρ_d is the density of firn/ice impermeable for water flow and $\rho_u = \rho_s + \Delta\rho$, where ρ_s is the density of fresh snow at the surface and $\Delta\rho$ is the contribution of the gravitational densification to the increase of a layer's density from ρ_s to ρ_d. The term $\frac{\rho_d - \rho_u}{\rho_d}$ can be seen as porosity of a seasonal snow layer by the end of ablation season corrected for future densification. Assuming ρ_d = 480 $kg\ m^{-3}$ and ρ_u = 800 $kg\ m^{-3}$ the term equals *ca* 0.4, suggesting that if in an annual cycle ablation exceeds 40% of accumulation then refreezing water can fill all the pores and additional water will contribute to the runoff. However, provided considerable spatial and temporal variability in ρ_d and ρ_u constraining of the parameters by empirical data is challenging.

Several methods was also suggested to estimate the rate of superimposed ice growth on top of a cold impermeable ice surface. Based on the results of a field campaign at the Barnes ice cap (Baffin Island, Canadian Arctic) and general theory of heat conduction and phase change Ward and Orvig (1953) suggested that the thickness of superimposed ice layer formed in a unit time is controlled by the amount of heat that can be conducted away from the accretion front. Implementation of this approach agreed plausibly with empirical evidences though relied heavily on assumptions regarding the initial ice temperature distribution, snow thickness and conduction rates. Later the approach was adopted with modifications and improvements by Woodward et al. (1997), Wakahama et al. (1976) and Bøggild (2007). Wright et al. (2007) suggested a method to predict the potential for superimposed ice formation as a function of the mean annual and winter air temperatures. The semi-empirical parameterization is based on measurements of the air and ice temperature at Midre Lovénbreen in Spitsbergen.

The approaches to include water refreezing in climatic glacier mass balance estimates presented above provide important first order estimates. Provided sufficient calibration using empirical data, they comply with the measured refreezing rates, however the simple approaches also have serious drawbacks. It is impossible to simultaneously take into account the principal processes governing the subsurface conditions. As a result a model tuned to perform well with respect to a dataset from a particular glacier or part thereof is likely to produce inaccurate results when forced by a dataset from a site within a different geographical setting.

A more complete description of the processes within the snow, firn and ice column is allowed by layered englacial models (e.g. Greuell and Konzelmann, 1994; Bassford et al., 2006; Van Pelt et al., 2012; Gascon et al., 2014). They simulate evolution of subsurface temperature, density and liquid water content in response to water infiltration and refreezing, heat conduction, gravitational settling and, possibly, other phenomena. An intercomparison of different approaches to quantification of refreezing was presented by Reijmer et al. (2012). It was shown that given reasonable tuning, simple parameterizations successfully reproduce the refreezing rates simulated for the Greenland ice sheet by advanced multilayer subsurface schemes within two different regional climate models (RCM). However, the spatial distribution of refreezing rates from parameterizations is significantly less consistent than between the two tested RCMs pointing to the lack of flexibility inherent to the simplistic approaches.

1.3 Zones on glaciers

The vast variability in characteristics of glaciers observed around the world explains the multiple attempts to classify them in several categories. For example, it is common to distinguish several morphological groups of glaciers depending on their respective shape (Ahlmann et al., 1933): ice sheets, valley glaciers, cirque glaciers, etc. However, in the perspective of the present thesis a classification dividing glaciers or parts thereof in different groups based on the subsurface conditions is more relevant.

Understanding of the fact that properties of the upper meters of snow and firn are controlled by the conditions at the surface made (Lagally, 1932) and (Ahlmann et al., 1933) independently from each other divide glaciers in three similar groups. Warm glaciers experience ample melt and have subfreezing temperature only in winter, transitional glaciers encounter moderate summer melt and at cold glaciers no melt is observed. The latter two groups have negative temperature down to some depth even in summer. This was followed by a debate on terms and definitions of units within a geophysical classification to be accepted (T.E.A., 1957). A detailed geophysical classification of glacier zones with a solid geophysical background was suggested by Shumskii (1955). The units of the classification were originally defined in terms of the dominant processes transforming snow to ice. These are, however, closely

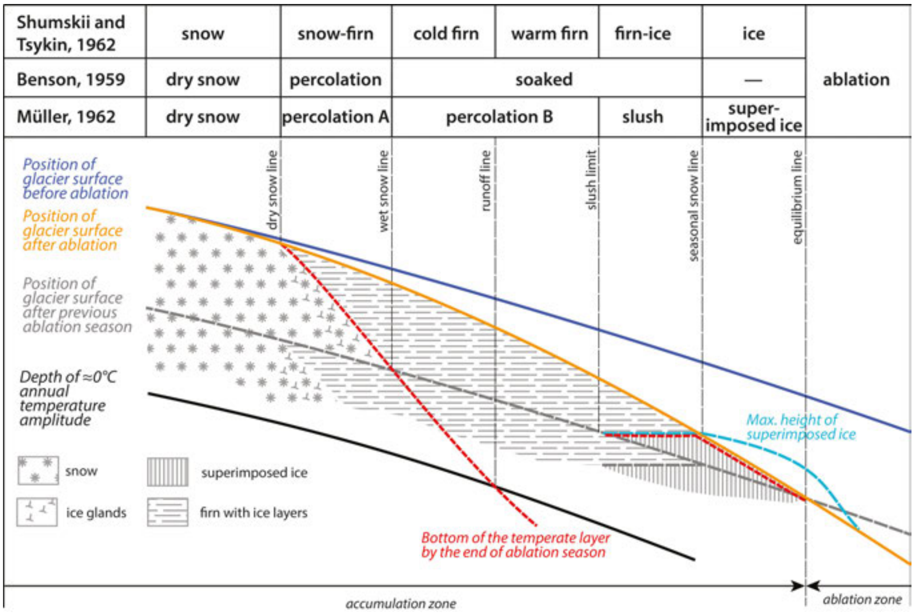

Figure 1.2. Geophysical classifications of glacier zones (facies) according to Shumskii and Tsykin (1962), Benson (1959) and Müller (1962).

linked with the subsurface stratigraphy, density, temperature and climate at the surface. The relevance of the classification for current glaciological research is explained by the fact that each unit is associated with a particular range of relations between three parameters characterizing the surface energy and mass fluxes. The parameters can be expressed in same units (for example, units of mass). Below W denotes the sum of melt and rain water. The amount of water that can be refrozen in the part of the subsurface profile that experiences annual temperature fluctuations (*ca* upper 10–12 *m*, referred below as ”active layer”) is expressed as RC and is mostly dependent on the surface energy flux during the cold season. The water mass that can be refrozen in pores of seasonal snow layer that remains by the end of ablation season is referred to as P. It is primarily controlled by the precipitation rate. It has to be noted that the zones are defined assuming stable climate and are thus an abstraction which is yet important for understanding the feedbacks existing between processes at and below the surface of glaciers even in a changing climate.

Below in this chapter the updated version the classification (Shumskii and Tsykin, 1962)) is reviewed starting from the highest zones and ending with the ablation area. The zones are correlated (Figure 1.2) with units from the classifications by Benson (1959) and Müller (1962). The two latter authors promoted a more descriptive approach to acknowledge the fact that strikingly different characteristics of the subsurface profile can be observed in different parts of a glacier.

Under cold and relatively precipitation-rich conditions melt rate is low and available liquid water entirely refreezes in the upper porous layer of snow and firn. This includes the snow, snow-firn and cold firn zones as arranged in the order of increasing melt rate. In the *snow zone* no melt is observed and it corresponds to the dry snow facies (zone) according of Benson (1959) and Müller (1962). In the *snow-firn zone* corresponding to the percolation facies (Benson, 1959) and percolation zone A (Müller, 1962) the melt water does not penetrate below the seasonal snow layer. In the *cold firn zone* at least a fraction of the active layer remains cold by the end of the melt season. A characteristic feature of the upper three zones is that the subsurface conditions allow to refreeze the entire mass of available liquid water and, consequently, runoff is absent. To ensure the latter refreezing should not be limited neither by the available refreezing capacity of active layer nor by pore space of the annual accumulation layer:

$$W < RC \wedge W < P.$$

At lower altitudes and latitudes the amount of available liquid water increases, while the precipitation rate usually drops. The part of a glacier where the following equation holds:

$$W \geq RC \wedge P \geq RC$$

is defined as the *warm firn zone*. Combined with the above described cold firn zone is corresponds to the percolation zone B according to Müller (1962). Here the active layer is not cooled enough during winter season to ensure refreezing of all the liquid water available. Water percolates deeper than the depth of zero annual amplitudes and warms the snow-firn column up to the melting point. After that the capillary forces within the snow/firn pack can still hold some additional water and prevent runoff. But if melting continues water will be evacuated through englacial and subglacial conduits and channels. In areas where runoff is restricted by shallow slope, water can accumulate over time forming vast slush fields referred to as perennial firn aquifers (Fountain, 1989; Forster et al., 2014; Miège et al., 2016; Christianson et al., 2015). It is, however, not obvious whether these features are self sustainable or are an attribute of abrupt climate changes. Ice in the warm firn zone is temperate unless the rate and pattern of ice flow from the higher and colder regions ensures a steady and sufficient supply of cold material.

In case the climatic conditions are dominated by a cold and long winter with limited precipitation, a considerable fraction of the pore space within the seasonal snow layer is filled by refrozen water. At the onset of melt water infiltrates through the snow pack and creates a slush layer on top of the impermeable ice surface formed during previous ablation period. Due to the steep temperature gradient in ice, caused by the low winter temperature, the slush layer gradually freezes forming a layer of superimposed ice. A fraction of melt water will inevitably run off on top of the ice layer by this evacuating a considerable amount of latent heat. By the end of ablation period all but a small fraction of the seasonal snow layer is either melted or filled by superimposed ice. The respective pore space will be filled by refreezing water during the early part of the next ablation season and thus snow undergoes transformation to ice within two years. The part of glacier where the above described processes are maintained belong to the *firn-ice zone* and the following relations have to be kept:

$$W \geq RC \wedge P \approx RC.$$

The zone corresponds to the slush zone according to Müller (1962). The conditions described by Benson (1959) as the soaked facies unite the domains of the cold firn, warm firn and the above described firn-ice zone.

If by the end of summer ablation removes the remains of seasonal snow and, possibly, also the upper part of newly formed superimposed ice, then bare ice is exposed to the surface by the beginning of winter. To ensure that the amount of precipitation should be relatively low, while severe conditions during long winter should ensured a significant refreezing capacity of the subsurface profile:

$$W \geq RC \wedge P \leq RC.$$

Such conditions are characteristic for the *ice zone* according to Shumskii and Tsykin (1962) and defined as the superimposed ice zone by Müller (1962). Here formation of ice takes one annual cycle.

At lower elevations the ablation becomes so intensive that the seasonal snow layer and, possibly, newly formed superimposed ice melt and contribute to the runoff. This results in negative net annual surface mass flux. The respective zone is recognized as the *ablation zone*. It has to be noted, though, that water refreezing is to be taken into account here too. The water refrozen during the summer season has to be melted again later, by this decreasing the cumulative ablation and runoff.

Above the glacier zones were described in the order that one possibly encounters if traveling downwards from the uppermost reaches of an abstract glacier. However, in most cases not all zones are to be found on every single glacier. The selection and order of zones at the glacier in question are a function of the local climatic conditions. The latter change not only with elevation, but also depending on latitude, level of continentality and over time. The general spatiotemporal pattern of change in glacier glacier zones is presented in Figure 1.3 adopted from (Shumskii, 1955; Shumskii and Tsykin, 1962).

Regions with maritime climate receive more precipitation, while winters are usually less severe. Here large parts of glaciers lie in the warm firn zone. Examples are glaciers in Iceland, Alps, Southern Alps in New Zealand, southern Norway and western Caucasus (Krenke, 1982). In more continental environments with large annual air temperature amplitudes and lower precipitation rates superimposed ice formation is more probable and firn-ice and ice zones are common. Such glaciers are found in the Svalbard(e. g. Storöyjökulen and Midtre Lovénbreen), Severnaya Zemlya, Canadian Arctic (e. g. Penny ice cap, Barnes ice cap), Western slope of the Greenland ice sheet, glaciers in northern Siberia (Byrranga, Verkhoyansk ranges) (Krenke, 1982). Changes in the boundaries between zones may be registered using methods of remote sensing (Wolken et al., 2009). Under conditions of climate change boundaries among zones are likely to be shifted and it is important that models describing internal accumulation are able to reflect this adequately.

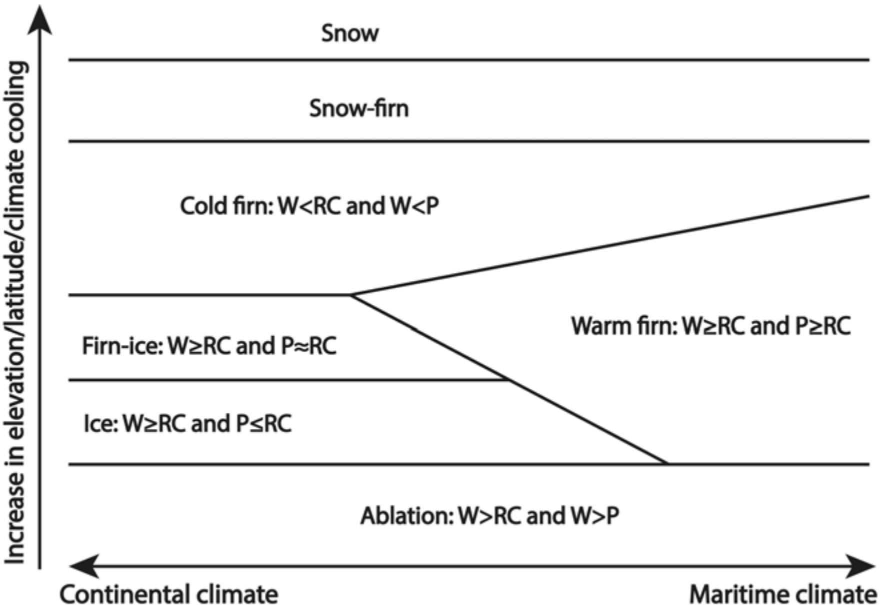

Figure 1.3. Geography of glacier zones. Adapted from (Shumskii and Tsykin, 1962).

1.4 Study site — Lomonosovfonna

The present thesis is to a large extent based on field data from Lomonosovfonna, a large accumulation field in the central part of Spitsbergen island, Svalbard (see Figure 1 in Paper I). In the present section a brief overview of the past glaciological studies at Lomonosovfonna is presented along with information on basic characteristics of local physical geography.

Lomonosovfonna is named after a Russian scientist and wrighter Mikhail V. Lomonosov (1711–1765). The vast glacier accumulation zone nourishes several outlet glaciers, of which most notable are Mittag-Lefflerbreen, Gruzdevbreen and Nordenkiöldbreen. The latter is called after a Swedish geologist and Arctic explorer Adolf Erik Nordenskiöld (1832–1901). It is one of the largest accumulation fields in Spitsbergen and is relatively easily accessible by a snow-scooter from Longyearbyen, the currently largest settlement on the island.

The latter fact makes Lomonosovfonna and Nordenkiöldbreen attractive for glaciological studies, which were started in summer 1965 by an expedition organized by the Institute of Geography of the Soviet Academy of Sciences (Zinger et al., 1966) when measurements of accumulation and ablation rates along with subsurface density and temperature observations were done. This was followed by a drilling campaign and subsurface temperature measurements in summer 1976 (Zagorodnov and Zotikov, 1980). Between 1991–1997 the Oslo university in cooperation with The Norwegian Water Resources and Energy Directorate (NVE) and the Norwegian Polar Institute (NPI) measured GPS positions of a stake profile along Nordenkiöldbreen and Lomonosovfonna yielding the ice flow speeds (Hagen et al., 2005).

Two deep ice cores were drilled at Lomonosovfonna in 1997 and 2009. The first one was the product of cooperation between the NPI, Universities of Lapland, Oslo, Tallin, Uppsala, Utrecht and other institutions. The core and the temperature measurements in the hole provided information on the chemical composition of ice (Isaksson et al., 2001; Pohjola et al., 2002b), evolution of the accumulation rate (Pohjola et al., 2002a) and air temperature (Van de Wal et al., 2002). The updated depth-time model constrained by the peaks in radioactivity and volcanic chemical signature (Kekonen et al., 2005) was later used to reconstruct the air temperature at Lomonosovfonna starting from the 8^{th} century AD (Divine et al., 2011). In 2009 another deep ice core was drilled in 2009 by a Norwegian-Swiss-Swedish expedition. Analysis resulted in detailed chemical record including nitrate and ammonium concentrations (Wendl et al., 2015; Vega et al., 2015).

Regular studies of ice dynamics and surface and subsurface mass and energy fluxes at Nordenkiöldbreen and Lomonosovfonna are done since 1999. The field campaigns with participators from Uppsala University, Utrech University and other institutions include mass balance measurements at a series of stakes, density and stratigraphy studies in shallow cores and snow pits, GPS measurements at stakes, radar profiling, etc. Since 2009 an automatic weather station is operated by the Utrecht University at ca 600 $m\ a.s.l.$

The field data used in the present thesis and associated methods are described in chapter 4. The study site is situated at 78.824° N, 17.432° E in the accumulation zone of Lomonosovfonna. The local altitude is $\approx$1200 $m\ a.s.l.$, while the equilibrium line at Nordenkiöldbreen was earlier estimated to be at ca 724 $m\ a.s.l.$ (Van Pelt et al., 2012). The glacier here is ca 192$\pm$5.1 m thick (Pettersson, 2009; Van Pelt et al., 2013) with a considerable layer of snow and firn in the upper part of the profile. According to Wendl (2014) in 2009 the firn thickness at the site was ca 20 m. Estimates of the rates of accumulation from repeated radar surveys (Pälli et al., 2002; Van Pelt et al., 2014) and ablation from modeled surface energy and mass fluxes (Van Pelt et al., 2012) yielded the values of 0.58–0.75 and 0.34 $m\ w.\ e.\ year^{-1}$.

Percolation and refreezing of melt water heavily influence the stratigraphy, density and temperature of the snow/firn pack at Lomonosovfonna. According to the subsurface measurements reported by Zinger et al. (1966) and Zagorodnov and Zotikov (1980) at least some parts of Lomonosovfonna belonged to the cold firn zone (see section 1.3 above for definitions). Our recent measurements (Paper I, Paper III) suggest that water refreezing warms the upper 10–12 m of the glacier profile, which is typical for the warm firn zone. This is well illustrated by Figure 1.4 showing the dynamics of the late-summer subsurface temperature at Lomonosovfonna during 1965–2016. This goes well in line with the recent finding regarding the regional climate warming rates Førland et al. (2011) and associated increase of the melt rates (Van Pelt et al., 2012, 2016; Østby et al., 2017). However, such empirical evidences as completely subfreezing temperature profile measured at 1250 $m\ a.s.l.$ in May 1997 (Van de Wal et al., 2002) and complex spatial variability of accumulation rate, affected by the wind drift (Van Pelt and Kohler, 2015), point to the possibly complex local pattern in the lateral change of subsurface conditions.

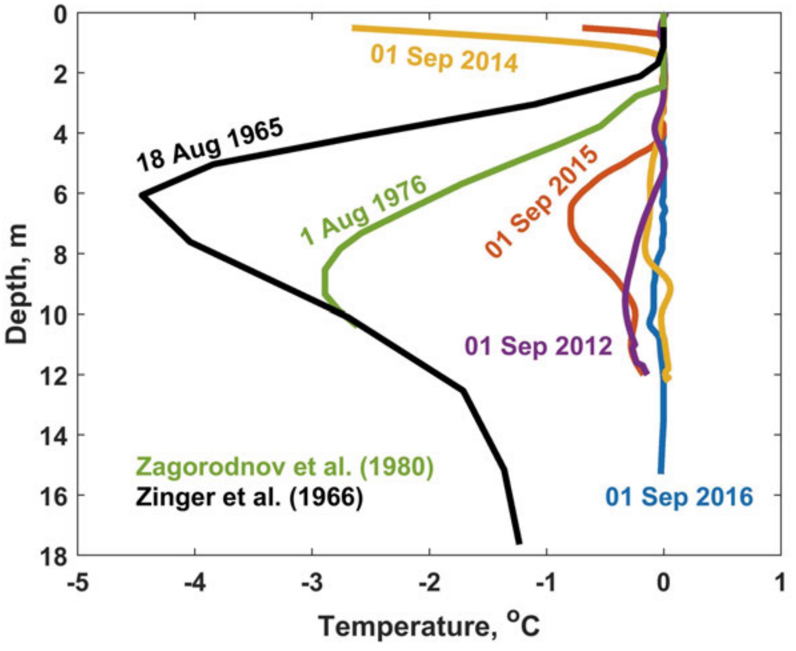

Figure 1.4. Change in late summer temperature distribution in snow and firn at Lomonosovfonna during 1965-2016.

2. Motivation

The primary motivation for taking into account the subsurface processes in glacier models is the expected more accurate estimation of the glacier runoff. Refreezing and retention in snow and firn result in a non-linear relation between water discharge from glaciers and the rates of liquid water supply at the surface (see sections 1.2.2 and 1.3 above) and thus deserve a careful description.

During the recent decades the dominant tendency of climate change is warming. According to Hartmann et al. (2013) the globally averaged Earth surface temperature has increased by ca $0.85°C$ during 1880–2012 and by ca $0.72°C$ during 1951–2012. In the Arctic the observed warming rates are significantly higher, which is largely attributed to the positive albedo-feedback between atmospheric warming and shrinking sea ice cover (e.g. Serreze and Barry, 2011). Increasing air temperature result in elevated melt rates, and overall retreat of glaciers (Vaughan et al., 2013).

The role of subsurface processes in the ongoing warming can be exemplified by the study from Pfeffer et al. (1991), according to which the forecast sea level rise caused by runoff from the Greenland ice sheet during the coming 150 years may be overestimated by as much as 5 cm if refreezing is not taken into account. However, as it was earlier mentioned, refreezing within a subsurface glacier profile is controlled by the temperature and pore space therein and is not infinite. The response of glacier-induced runoff to a climate perturbation is defined by the interplay between the characteristics of the purturbation and the subsurface conditions in question.

For instance, is has been shown (Noël et al., 2017) that at the glaciers and ice caps on Greenland climate warming since 1958 has first resulted in increase of the refreezing rates, but after 1997 (± 5 y) this was followed by a decline in that term of the climatic mass balance. At the same time net refreezing below the surface of the contiguous but higher elevated and colder Greenland ice sheet has been growing steadily with the post-1997 tendency being even somewhat higher. The difference between the two types of glaciation on Greenland in response to the climate change is also seen in runoff dynamics after the 1997 refreezing tipping point. The corresponding values for the ice sheet rise in pace with the melt, while discharge from glaciers and ice caps climbs up 65% faster than the simultaneous melt water production. However, it was also shown that in case of continued climatic warming a similar evolution of the melt-runoff dynamics can be expected at the Greenland ice sheet (van Angelen et al., 2013).

Empirical evidences from different regions point to the fact that snow and firn at glaciers can exhibit contrasting response to climatic changes not only due to differences in elevation. For example, the lower part of the accumulation zone on the western and eastern slopes of the Greenland ice sheet show diverging behavior of the respective subsurface profiles. One of the notable recent discoveries in Greenland is the perennial firn aquifers (Forster et al., 2014) found mostly at the south-eastern part of the ice sheet (Miège et al., 2016). Gentle slopes along with ample precipitation and ablation do not encourage effective evacuation of melt water downslope (Munneke et al., 2014). According to the field observation on the western slope of Greenland, thick ice layers are forming in the lower accumulation area, yielding the underlying firn largely inaccessible for melt water refreezing and retention (Machguth et al., 2016). Similar processes have also been reported from the Devon ice cap in the Canadian Arctic by Gascon et al. (2013). It is worth noting that the eastern slopes of Greenland receive significantly more precipitation than both western Greenland and the Canadian Arctic (Miège et al., 2013; Koenig et al., 2016).

The above described examples of changes of the subsurface properties at glaciers can be seen in the perspective of spatiotemporal dynamics of glacier zones following the scenario characteristic of either maritime or continental climate (right and left sides on Figure 1.3). For the glaciers of the study region (Svalbard) Van Pelt et al. (2016) showed that, in case of continued warming, cold firn zones are likely to be replaced by warm firn zones in a combination with a narrow band of ice zone nourished by superimposed ice. However, it is known that superimposed ice plays a very important role at some glaciers, for example: Midtre Lovénbreen (Wadham and Nuttall, 2002; Wright et al., 2005), Kongsvegen (Obleitner and Lehning, 2004) or Storöyjökulen (Jonsson and Hansson, 1990), entirely nourished by superimposed ice.

A number of approaches was suggested to reduce the uncertainty in assessments of the climatic glacier mass balance by taking into account the processes in snow and firn, as was reviewed in chapter 1.2.2. Advanced multilayer models describing the subsurface processes have been shown to produce prominent results (e.g. Fettweis, 2007; Lenaerts et al., 2012; Steger et al., 2017). However, these one-dimensional models rely on a number of assumptions regarding and parameterizations of inherently multidimensional subgrid processes that either contradict empirical evidences or are poorly constrained by field data. It is worth noting that a model never gets better than the data used for forcing and the physics driving the simulations. Thus to be able to describe the complex response of glaciers to a changing climate, demonstrated above, it is essential to increase the level of understanding of process governing the mass and energy exchange in snow and firn.

A major challenge is a formalized description of the sophisticated rigid structure of snow, firn and ice caused by the fact that in natural conditions ice is closer to the pressure melting point than most other commonly met media.

Thus porous snow and firn are subject to rapid mechanical and thermodynamic changes both in space and time. This was illustrated by e.g. Schneebeli and Sokratov (2004) using 3D tomography of centimeter-scale snow samples. Sturm and Benson (2004) reviewed the phenomena of heterogeneity for the seasonal snow pack at the scales from 10 to 10^5 *m* and Dunse et al. (2008) for a $100\times100\times10$ *m* volume of firn. As a result the physical formulations describing processes below surfaces of glaciers are challenging to formulate and associated subgrid parameterizations based on scarce empirical data are often not accurate.

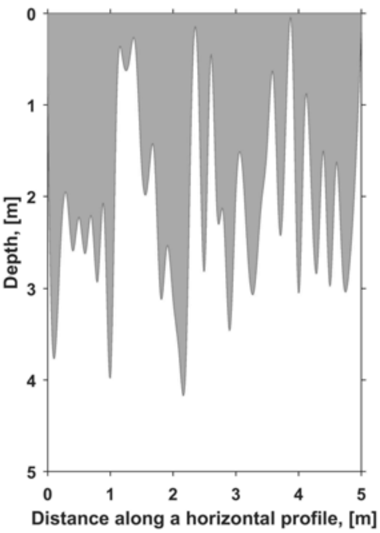

Figure 2.1. A schematic representation of preferential water flow in a snow and firn pack - shaded area shows the wet part of an imaginative snow pit wall. The scales and the shape of the ”fingering front” are chosen arbitrarily for illustrative purposes.

One of such poorly understood processes is water flow through snow and firn, although the associated fluxes of mass and latent heat are a very efficient way to rise the subsurface temperature and density. It is commonly assumed that the horizontal gradients in vertical propagation of water are absent. However, multiple evidences of high lateral inhomogeneity of the water infiltration have been reported (e.g. Marsh and Woo, 1984; Schneebeli, 1995; Bøggild, 2000; Waldner et al., 2004; Campbell et al., 2006). Thus water can be found much deeper than predicted by simulations relying on the wetting front hypothesis (Figure 2.1) and it is impossible to account for the associated refreezing-induced changes in temperature and density. The uncertainties in the latter two parameters will then propagate further in descriptions of other processes.

For example the rate of downwards propagation of the winter cold wave is to a large extent controlled by refreezing of the water stored in the snow/firn pores. However, the above described effects of preferential water flow result in significant volumes of dry snow and firn left above the fingering front (e.g. Techel and Pielmeier, 2011). Thus parameterizations of capillary water retention properties of snow/firn derived by fully saturating samples and then letting water drain (Coléou and Lesaffre, 1998) may not be directly applicable for describing thick snow/firn packs. This is particularly relevant at the onset of ablation and in areas with generally low melt rates, as then the effect of preferential flow can be expected to be most pronounced.

Conductive heat exchange dominates subsurface energy fluxes during a considerable part of a year and is the only process resulting in increase of the snow/firn/ice refreezing capacity. It is driven by the temperature gradient and is controlled by the effective thermal conductivity (k_{eff}), which is often parameterized as a function of density (Sturm et al., 1997). However, there is a vast spread across different parameterizations (Sturm et al., 1997), pointing to the fact that density is not the only argument controlling k_{eff}. Moreover, few k_{eff} estimates are published for dense snow and firn samples, which further increases the uncertainty range in simulations involving conductive heat exchange.

Above we have named the aspects of snow/firn studies motivating the present thesis. All of them are united by the ambition of obtaining an accurate description of the interactions between constantly changing climatic conditions and melt-induced runoff from glaciers. There are, however, other circumstances motivating glaciological studies of snow and firn. For example, traveling on glaciers (e.g. Rivera et al., 2005) and environmental pollution (e.g. Colgan et al., 2016) are important practical implications of snow and firn studies.

3. Aim and objectives

The present thesis aims to contribute to the increase of our understanding of the processes involved in mass and energy exchange in snow and firn at glaciers with a particular focus on the accumulation zones.

This is dictated by the framework of the project named "Stability and Variations of Arctic Land Ice" (SVALI), in which the present study was included as one of the work-packages. Together with other two projects SVALI belonged to the sub-programme "Interaction between Climate Change and the Cryosphere" of the Top-level Research Initiative, a major Nordic collaborative venture for studies of climate, energy and the environment. The work was also included in the research done within a project funded by the Swedish Research Council (Vetenskapsrådet) and focused on similar agenda.

Below we first list the objectives formulated before the start of the project on the basis of the relevant SVALI deliverables and general understanding of the task:

1. review the existing approaches to quantification of refreezing and published refreezing rates from Svalbard glaciers
2. identify potential areas of advance, provided the collaborative possibilities and the research infrastructure available
3. conduct field observations and measurements of subsurface mass and energy fluxes at Lomonosovfonna
4. complement a commonly used refreezing scheme with a suitable water routing model

The beauty and danger of research is that at the start of a study it is seldom obvious how exactly the scientific task is to be solved. That particularly applies to studies dealing with such poorly controlled and challenging environments as glaciers. Thus below we also state the objectives of the thesis once suggested to the scrutiny of the general and professional audience:

1. present the studies reported in the four papers forming the core of the thesis
2. using the empirical data illustrate and quantify the structural complexity of the snow, firn and ice at Lomonosovfonna and the associated spatio-temporal heterogeneity of the subsurface mass and energy fluxes
3. exemplify the fact that the sophisticated nature of subsurface processes has also patterns that can be both observed and described, provided that sufficient empirical evidences are available and adequate simulation approaches are used

4. Field data and methods

The present thesis largely relies on empirical data collected at the Lomonosovfonna ice field during six field campaigns in April 2012–2017. The author participated in five expeditions skipping the field operations in 2013. The data collected in the field was used to constrain get an insight into the subsurface processes and to constrain several models describing the energy and mass fluxes at and below the surface of the ice field.

4.1 Surface mass and energy fluxes

Processes in firn are largely controlled by the ablation and accumulation rates at the surface. To constrain the upper boundary conditions for the subsurface scheme applied in Paper II, data from a mass balance stake and two automatic weather stations (AWS) installed at the study site was used.

The stake named ”S11” is found at 78.8114°N 17.4540°E, 1147 $m\,a.s.l.$ It is the highest in the series of stakes used by the Uppsala and Utrecht Universities for monitoring of the surface mass balance at Nordenskiöldbreen and Lomonosovfonna. The distances between the top of the stakes and the surface of snow/firn/ice is measured annually during the field campaigns in March–April. Displacements of the stakes in the horizontal and vertical directions are continually tracked using GPS units mounted on the top.

The two AWS installed at Lomonosovfonna included the following set of instruments: air temperature and relative humidity sensor, a sonic ranger for tracking the changes in relative surface height and radiometer. The latter consisted of the $SW\downarrow$, $SW\uparrow$, $LW\downarrow$ and $LW\uparrow$ sensors. The first AWS was mounted in April 2013 by Christian Zdanowicz and Rickard Pettersson with the help of Veijo Pohjola. It was operational until July 2013, when the logger box was flooded by water resulting in hardware malfunction. The AWS was excavated from the layers of snow and rime

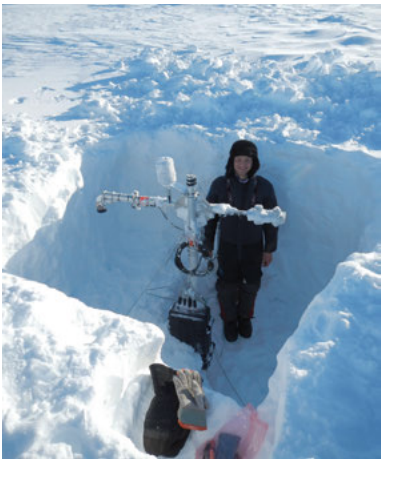

Figure 4.1. AWS installed in April 2013 after a year of accumulation and some hours of digging.

accumulation next year in April 2014. It was replaced by a new installation, which was functional during the following year. However, extensive accumulation of rime and snow had a dramatic effect on the quality of readings from the radiometer and sonic ranger.

4.2 Subsurface density

Four shallow firn cores were drilled at Lomonosovfonna in April 2012–2015 using a Kovacs corer with the internal diameter of 9.5 *cm*. Density measurements and stratigraphy observations (see chapter 4.3.1 below) of the cores were done in the cold laboratories of the University Centre in Svalbard (UNIS) and of the Norwegian Polar Institute (NPI) by the author (2012, 2014 and 2015) and Carmen Vega (2013).

To calculate the density of core samples they were weighted and their volume was approximated from measured geometrical dimensions. In case the shape of a sample was irregular, it was cut by a band saw to obtain a cylinder or a rectangular prism. Volume of these convex shapes is calculated using the measurements of diameter and length and of lengths of three sides. In 2012 the cores were cut in The resulting data on subsurface density distribution was employed in all four papers presented in the thesis.

4.3 Subsurface stratigraphy

To obtain information about layering in the subsurface profile at the field site three different methods were used. They are briefly described in the following sections. Comparison of results from the methods formed an important part of Paper I.

4.3.1 Cores

Stratigraphy of the four firn cores drilled at Lomonosovfonna was described and documented using a photo camera. Cores examined in the cold laboratory of the NPI were viewed in the transmitted light from a fluorescent lamp mounted under a transparent table. Originally the stratigraphic classification proposed by Pohjola et al. (2002b) was used. Later during the data analysis this detailed description was simplified by distinguishing only between solid ice and porous icy media.

Parts of the core were lost in the drilling operations and during handling and transportation. The interrupted stratigraphic sequence of the core drilled in 2012 was reconstructed with the help of continuous video records from the core hole. When such correlations were not possible it was assumed that the core pieces rather belong to the upper part of the profile than the lower, since the coring dogs can slide some distance along the sample.

4.3.2 Borehole videos

Stratigraphic observations of snow and firn in boreholes was done using specialized camera equipment provided by UNIS. The used camera unit is a compact device consisting of an image sensor and a light source. It was lowered in 5.5 *cm* wide boreholes drilled using a Kovacs auger. The current depth of the camera was controlled by means of a digital cable counter devise linked to a video processing unit, which overlaid depth marks on the main video signal. The resulting data was recorded on a computer placed in a mobile shelter at the surface. During the first field campaign in April 2012 thirteen boreholes were filmed by the author of the thesis and Horst Machguth. In 2013 five holes were filmed by Dorothée Vallot and Rickard Pettersson. In 2014 the thesis author filmed four holes with the help of Dorothée Vallot.

Analysis of the videos was a laborious and tedious undertaking that consisted of watching the video records and making notes regarding the apparent changes in structure of the subsurface layers. It was possible to see ice layers as thin as a few millimeters as well as transitions between courser- and finer-grained snow/firn. However, the initial more detailed subsurface descriptions were then reconsidered and distinction only between porous snow/firn and solid ice was made.

An attempt to optimize the process of ice lens detection was also made. The approach was based on the difference in optical properties between solid ice and the porous snow and firn. The latter has lower albedo and thus the associated pixels in the video frames can be expected to appear darker than the surrounding porous material. The videos were broken in sequences of individual frames and the brightness of the same fragment on each ("control window") was calculated. To correlate each brightness value with a particular depth reading the fragment of the corresponding frame containing depth information was analyzed. For that an optical character recognition routine was implemented using the Neural Networks toolbox in Matlab software. The algorithm for ice lens detection returned ambiguous results. Most of the ice lenses observed in the videos could be matched with local lows in brightness of the control window. The main problem was that multiple brightness minima could not be correlated in depth with any of the ice lenses subjectively interpreted from the videos. That was caused by multiple factors. Firstly, the used equipment did not allow to manually control the sensitivity of the camera digital sensor. Secondly, the overall quality of the videos was largely affected by drilling chips sticking to the borehole walls and fogging of the camera lens as it was lowered from the colder upper part of the profile towards the deeper and warmer firn layers.

4.3.3 Ground penetrating radar

Stratigraphy of the subsurface layering at Lomonosovfonna was also studied using the electromagnetic properties of snow/firn/ice as a proxy. For that Ground Penetrating Radar (GPR) data was collected during the field campaigns in April 2012–2014.

Radars (from radio detection and ranging) have long been used for deriving information about thickness and properties of different objects (Jol, 2009) and are widely applied in glaciology (e.g. Siegert, 1999). The propagation of electromagnetic waves through media is determined by the electrical permittivity (ε) and conductivity (σ) as well as by the characteristics of the equipment used to send and receive the waves. Reflection of an electromagnetic wave occurs at sharp gradients in ε and/or σ (Hubbard and Glasser, 2005), most often caused by spatial changes in relative amount of solid ice, air and water, constituting bulk glacier profile and having very different electromagnetic properties. Thus GPR reflections can be associated with such phenomena as ice layers or volumes within snow/firn, boundaries between annual layers marked by the layers or boundaries between subfreezing and temperate volumes as the latter usually have non-zero water content.

In the present thesis data from a Malå ProExpulse radar system with antennas having the center frequency of 800 MHz operated by Rickard Pettersson and Katrin Lindbäck is used. Dense series of profiles covering the area of ca 10×10 m and separated by 0.2 m were done. After application of the post-processing routines (see Paper 1 for details) this allowed to derive three-dimensional descriptions of the subsurface structures at the field site.

4.4 Subsurface temperature

Temperature of the subsurface snow, firn and ice layers at glaciers is a sensitive parameter that exhibits a pronounced response to the surface temperature fluctuations and to the release of latent heat accompanying liquid water refreezing. It is thus a crucial proxy for describing the subsurface energy and mass fluxes and its measurements make an important part of the present thesis.

Measurements of temperature evolution in the snow and firn profile at Lomonosovfonna were done with the help of thermistors, which are resistors precisely manufactured from a material that changes resistivity in response to temperature. Thus temperature estimates are based on measurements of resistance, which characterizes the difficulty of electrical charges to pass through a conductor. Resistance an electrical circuit section (R) is commonly measured by applying to it a known electrical current (I) and measuring the resulting voltage drop (V). Then resistance is calculated using the Ohm's law as:

$$R = \frac{V}{I}$$

Resistance measurements used in the thesis were done using data loggers manufactured by Campbell Scientific (model CR10X). These devices are able of applying precisely measured voltage to several of its ports and also to precisely measure the voltage that is applied to its other ports. To take advantage of that a simple electrical circuit was constructed (Figure 4.2). The two wires of every thermistor were connected to the excitation (EX) and analogue input (CH) ports on the data logger. The reference temperature-stable resistor with known resistance R_r was joined in series with the sensor by connecting its two leads to the same analogue input port (CH) and analogue ground port (AG). The logger was programmed to apply a precisely measured excitation voltage drop V_{ex} between the EX and AG ports and then measure the voltage drop V_m over the reference resistor (between CH and AG). Since the thermistor and reference resistor are connected in series, the current (I) is constant in the entire circuit. Writing out Ohm's law for the parts of circuit from EX to AG and from CH to AG yields:

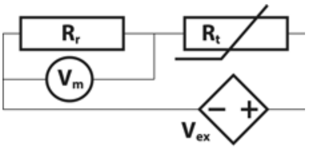

Figure 4.2. Electric circuit for measuring resistance of a thermistor

$$\frac{V_m}{R_r} = \frac{V_{ex}}{R_r + R_t}$$

which after rearrangement returns the sought thermistor resistance:

$$R_t = R_r \cdot \frac{V_{ex}}{V_m} - R_r$$

After that resistances calculated from voltage measurements were further converted to temperature values following the information provided by the sensor manufacturer. Functions relating the resistance and temperature of a thermistor are often based on the Steinhart-Hart equation (Steinhart and Hart, 1968).

Multiple thermistors were arranged in thermistor-strings (T-strings), by soldering individual sensors on leads within a multi-leaded cable (Figure 4.3) and isolating the contacts. The cable was then covered by self-amalgamating tape and a thicker shrink tube for mechanical and water protection. The number of temperature sensors by far exceeded the number of channels in the applied data loggers and relay multiplexers from Campbell scientific were used to successively connect several thermistors to one channel.

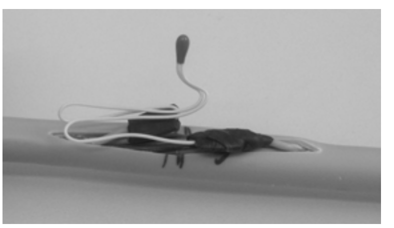

Figure 4.3. Thermistor soldered into a wire of a multi-leaded cable. Black patches are pieces of shrink tube isolating the contacts.

Nine T-strings were manufactured by the author in each of the years from 2012 to 2014. They were installed in boreholes drilled by a Kovacs auger in a rectangular 3×3 pattern with a 3 m spacing between nodes. The holes were then backfilled using loose snow and drilling chips. In April 2013 the strings were installed by Christian Zdanowicz, while in the other two years that was done by the author. One difficulty encountered in the field was that the PVC coating of the cable became very stiff when cooled down to -10...-20$^{\circ}C$ and difficult to straighten. Consequently, the real separation between sensors tended to be smaller than the designed one, necessitating additional efforts to straighten the cables. In 2014 T-strings were warmed up in a box using a portable heater before uncoiling.

In 2015 the design of the T-strings was updated to reduce the effect: the multi-leaded cable was put into 2 m long rigid plastic tubes which were fixed at certain positions by cable glands attached to the sides. The sensors were fed through holes in the tubes and fixed with predefined intervals on their outer sides. One such T-string was manufactured and installed in April 2015.

5. Summary of papers

5.1 Paper I

The aim of this study was to derive a detailed description of the spatial and temporal variability in subsurface stratigraphy at Lomonosovfonna. For that three different methods were applied for ca $10\times10\times10$ m volumes of firn during three field campaigns in April 2012–2014. Shallow cores and borehole inspections using a video camera allowed to observe the subsurface layers using their optical characteristics, while GPR surveys rely on the dielectric properties of the media as a proxy.

Occurrence of the ice volumes found in individual boreholes had a pronounced lateral variability. That was attributed to the effects of horizontally inhomogeneous water infiltration pattern. At the same time the stratigraphy signal averaged over multiple holes was much more consistent: ice layers in firn tended to be concentrated in several preferential horizons. The same applied for the GPR surveys. Reflectors in single profiles seem chaotic and could not be correlated with ice layers found in the boreholes drilled within less than 0.1 m from some of the radar traces. However, the average of all profiles collected in the same field campaign shows distinct laterally consistent internal reflecting horizons. The latter could be correlated in depth with the above mentioned horizons of preferential ice layer formation and several of them were consistently seen in data from all three years. That along with earlier published results allowed to interpret them as buried summer surfaces from previous years.

Thus the main finding of the paper is that the lateral consistency in descriptions of subsurface stratigraphy is only seen after extensive averaging of data collected in points.

5.2 Paper II

This paper is based on comparison of the measured and simulated evolution of subsurface temperature during three summer seasons (2012–2014). Measurements were made by 27 (9 each year) continuously scanned T-strings. For simulations we used a one-dimensional multilayer snow/firn/ice scheme coupled to a surface mass and energy balance model forced by the output of the regional climate model Weather Research and Forecast (WRF).

Temperature of subfreezing snow and firn is very sensitive to introduction of liquid water. Thus the data shows:

- a high lateral inhomogeneity of water infiltration;
- high rate of warming of the subsurface profile after the onset of melt;
- that by the end of the three melt seasons the upper 10 to 12 *m* of the subsurface profile at Lomonosovfonna were temperate.

The two latter findings could not be reproduced using the conventional configuration of the subsurface scheme, allowing water to enter each next layer only after the refreezing and liquid water holding capacities of the current one are exhausted.

Motivated by the evidences of preferential water flow at the site, originating both from stratigraphy (Paper I) and subsurface temperature data, a routine describing the process was implemented in the snow/firn/ice model. The suggested approach is to first distribute the water generated at the surface at each time step among the subsurface layers. Water distribution along the profile is governed by percolation functions decreasing with depth. After that the conventional water infiltration routine is executed. Different patterns and percolation depths (z_{lim}) were explored.

As water is allowed to go deeper (by increasing z_{lim}) the root mean square differences between the simulated and measured temperature (*RMSD*) first show a steady decrease, then at some point a minimal *RMSD* value is reached, which is followed by an increase. This suggests that for each percolation pattern there is an optimal percolation depth resulting in most accurate temperature simulation during the summer period. The corresponding *RMSD* values are approximately two times lower compared to the results from the default water infiltration scheme. Mechanism of the suggested approach to description of the preferential water flow is based on underestimation of the potential of capillary forces to retain water against gravity as for each layer it is the second condition to be satisfied after refreezing capacity is eliminated. It was also shown that a reasonable increase in surface melt rate, that would not ruin the correlations between simulated surface energy balance parameters and AWS data, did not result in any significant drop of the *RMSD*.

The main conclusion from the paper is that while being an important subsurface process, preferential water flow in snow and firn can be taken into account within one-dimensional layered models using the suggested approach relying on percolation functions. Further development of the latter will, however, require extensive field measurements.

5.3 Paper III

This study aims to reconstruct the effective thermal conductivities in snow and firn at Lomonosovfonna based on the measurements of subsurface temperature evolution. If the upper part of a glacier profile has subfreezing temperature then its thermodynamics is defined by the conductive heat flux. Here it is described by the one-dimensional equation:

$$C\,\rho\,\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}(k_{eff}\frac{\partial T}{\partial z}), \tag{5.1}$$

where time t, depth z, density ρ, and temperature T are measured using shallow cores and T-strings and the specific heat capacity C is approximated from measured T values (Cuffey and Paterson, 2010). Thus, the only poorly constrained parameter in the equation is the effective thermal conductivity k_{eff} sought after.

Three subsurface temperature datasets with no values exceeding -2°C (spring 2014, fall-winter 2014, spring 2015) provided the boundary conditions driving the conduction model based on Equation 5.1. For each of the three model domains 2 to 4 ca 2 m thick layers with invariable k_{eff} are distinguished. The corresponding values are adjusted to minimize the misfit between the simulated and measured subsurface temperatures using an optimization routine. Additional numerical experiments were done to explore the potential uncertainties in the estimates of the optimized k_{eff} values that may be caused by the errors in employed empirical data.

The optimized k_{eff} values increase with depth and vary from 0.6 to 1.3 $J\ (s\ m\ K)^{-1}$, while being consistently and sometimes significantly higher than results from commonly used density-based parameterizations by Sturm et al. (1997) and Calonne et al. (2011). The latter finding suggests an underestimation of the effectiveness of the conductive heat exchange by subsurface models relying on the parameterizations.

5.4 Paper IV

The study also utilizes subsurface temperature data from Lomonosovfonna and largely relies on the k_{eff} estimates from Paper III. Here a new method for assessment of firn water content is suggested and applied for the subsurface profile at Lomonosovfonna.

The governing hypothesis for the study is that the evolution of subsurface temperature during the cold part of a year is defined by two processes: refreezing of water left in pores after the ablation season and conductive heat flux towards the cold surface. As a result of these processes the freezing front penetrates downwards over time, which is a phenomena that is easily interpreted from measured temperature evolution. It was thus suggested that the firn water content can be quantified from estimates of the effectiveness of conductive heat exchange and analysis of the freezing front propagation rates. Implementation of the approach was based on subsurface temperature evolution measured at Lomonosovfonna in September 2014–January 2015.

The resulting volumetric fractions of liquid water contained in firn pores after the melt season in 2014 are ca 1–2.5 $vol.\%$ for the upper part of the profile (above 3 *m* referenced to the glacier surface in April 2015) and less than 0.5 $vol.\%$ below. These values are significantly lower than results from the density-based parameterization (Schneider and Jansson, 2004) widely applied in multi-layered subsurface schemes. This, along with the prominent variability of the water content estimates inferred from different T-strings, was interpreted as an effect of preferential water flow that leaves ”pockets” of dry firn behind the fingering front.

6. Outlook

The observations and measurements done at Lomonosovfonna in the framework of the present thesis provide further evidences of the fact that at the scale of meters the structure of snow, firn and ice in accumulation zones of glaciers can be very sophisticated and hardly predictable from point measurements. A consequence of that are the difficulties in formalization of such properties and processes in subsurface simulation domains as: metamorphism of snow/firn grains, gravitational settling, conductive heat flux, infiltration of water and its subsequent refreezing, retention and runoff.

It is thus important to develop techniques suitable for detailed descriptions of the snow, firn and ice properties and their change in time. While direct evidences from cores and boreholes can not provide the necessary frequency of observations neither in time nor in space, they can be used to mitigate the shortcomings of the radar surveys by providing indispensable data for validation and calibration. GPR surveys are non-destructive, efficient in covering large volumes of subsurface material by measurements, can be operated autonomously and thus provide prominent time coverage. Finally GPR surveys rely on a solid physical background – the difference in dielectric properties between constituents of snow/firn: solid ice, water and air.

In Paper I it was shown that after extensive lateral averaging of the stratigraphy signals from the GPR and borehole camera surveys, the two methods returned patterns that are correlated in depth. This suggests that reflecting horizons in firn are associated with the ice layers. With the available GPR data it was not possible to go into finer details, but it would be an interesting and useful experiment to employ a high-resolution radar survey for characterization of the shape and size of ice volumes in a subfreezing snow/firn pack and of wet pockets with high water content in a temperate one.

The complex spatial structure of the porous ice media is to a large extent created by the sophisticated pattern of water flow and refreezing. In Papers II and IV it was shown that subsurface temperature is a very useful parameter for monitoring of the spatial and temporal pattern in water flow and refreezing events. Measurements using T-strings are cheap and relatively straightforward to interpret (compared to, for example, radar data), which favors their wide application for studies of mass and energy fluxes in the accumulation zones of glaciers.

Preferential water flow directly influenced the results of Papers I, II and IV. The process has important implications for the characteristics of a subsurface column by effectively transporting water into deep layers while not entering the snow and firn in between the flow fingers and sub-horizontal water impeding horizons. Primary consequence of that is fast elimination of subfreezing temperature in the profile by means of latent heat release and conductive heat fluxes. The latter can be directed not only vertically, but also horizontally, which needs to be taken into account, provided the reported anisotropy in the effective thermal conductivity of snow (Calonne et al., 2011; Riche and Schneebeli, 2013). By effectively warming deep firn layers preferential flow facilitates build up of a temperate firn and eventually ice column. Thus a model with an accurate description of the process is more likely to correctly reproduce the warm firn zone conditions at a site with the characteristic runoff regime.

The process of water transport also deserves a careful examination at the scale of kilometers. The relation between dynamics of water supply at the surface of glacier and runoff can experience a dramatic influence from such earlier mentioned features (see sections 1.3 and 2) as the vast perennial firn aquifers (PFA) and massive ice layers in the upper part of the firn pack. Both phenomena have been observed in the lower accumulation zones, however, the former are located in relatively maritime climates (Fountain, 1989; Forster et al., 2014; Christianson et al., 2015; Miège et al., 2016), while the latter seem to be restricted to regions with cold winters and a low accumulation rate (Gascon et al., 2013; Machguth et al., 2016).

This pattern in geography of the PFAs and massive ice layers in firn can be seen in a broader perspective of the concept of glacier zone migration both in space and time. As was stated in chapter 1.3 above, the temperature, density and stratigraphy of the snow/firn/ice profile are to a large extent controlled by the relation between three climate-controlled parameters: the annual amount of accumulation and ablation along with the effectiveness of surface cooling during winter. Modern regional climate models coupled to subsurface schemes (e.g. Reijmer et al., 2012; Van Pelt et al., 2016; Østby et al., 2017) open exciting perspectives for exploration of the feedbacks between processes at and below the surfaces of glaciers.

Firstly, it is important to validate the output of the simulated subsurface conditions against empirical data. That can be done by comparing the modeled and measured vertical distributions of temperature, density and stratigraphy. Secondly, the internally consistent simulation products provide valuable insights into the spatial and temporal dynamics of the firn conditions. For example they can be used to delineate glacier zones and by this locate the regions where certain firn properties are to be expected. Thirdly, it might be possible to draw conclusions regarding the rate and pattern of change in accumulation and melt rates, harshness and length of winters from comparisons between modern and past evidences from the snow/firn/ice columns.

7. Sammanfattning på svenska

Översättning från engelska gjord av Veijo Pohjola
Glaciärer täcker ca 10% av planeten jordens yta, där du främst hittas på höga höjder och breddgrader. De är viktiga komponenter i klimatsystemet på grund av de många återkopplingar som länkar dem till atmosfär, hydrosfär och det omgivande landskapet. Glaciärernas kalla och sluttande ytor påverkar den atmosfäriska cirkulationen, vilket i sin tur påverkar villkoren för glaciärens massbalans. Glaciärerna används som klimatarkiv för förståelse av tidigare klimatförändringar. Glaciärerna är den största färskvattenreservoaren på vår planet och smältning och avrinningen from dem påverkar den globala havsnivån och med detta det globala samhället. Dessutom påverkas de miljöer som ligger i anslutning till glaciala avrinningsområden. Att förstå avrinning från glaciärer beror till viss del av hur smältvattnet dräneras och lagras i glaciärernas porösa lager av firn. För att förstå hur mycket smältvatten som rinner ut ur en smältande glaciär pga av en klimatförändring måste vi förstå mass- och energiförhållandena i glaciärernas firnområden.

Denna avhandling fokuserar på processerna för mass- och energiutbyte i glaciärernas firnområden, som till stor del drivs av klimatdrivna energiutbytesprocesser vid firnytan. Resultaten i avhandlingen bygger i stor utsträckning på empiriska data från isfältet Lomonosovfonna, Svalbard, och har samlats in under fältkampanjer mellan 2012–2017.

Observationer av firndensitet och stratigrafi från iskärnor, videoupptagningar från borrhål och radarundersökningar gav en detaljerad beskrivning av snö- och firnlagren. Förekomsten av islinser som hittades i enskilda borrhål var hade en hög lateral variabilitet. Samtidigt var stratigrafin i genomsnitt mer konsekvent i den större skalan: islinserna tenderade att koncentreras i flera preferenshorisonter som återkom i flera av borrhålen. Detsamma gällde för isradarmätningarna. Reflektorer i enstaka profiler hade ett sporadiskt mönster och kunde inte korreleras med enstaka islager från borrhålsobservationerna. Medelvärdet av alla isradarprofiler som samlats in visar emellertid på tydliga horisonter som återkommer med jämna mellanrum i profilerna. De senare kunde korreleras med ishorisonterna från videoupptagningarna. Flera av dessa horisonter observerades konsekvent i data från alla de tre år som studien omfattade. Dessa lager tolkades som begravda sommarytor från tidigare år. Ett delfynd av avhandlingsarbetet är att firnstratigrafin endast kan visas efter areellt omfattande observationer som ett medelvärde den mängd punktdata som samlats in. Det är inte statistiskt möjligt att skapa en areellt täckande stratigrafi bara från ett fåtal punkter, som tex från en enda iskärna.

Den firntemperaturdata som samlats in med hjälp av ett flertal vertikala termistorsträngar gav insikter om de termodynamiska processer som opererar i den porösa firnen. Temperaturmätningar i firn och snö är känsliga för vattenflöden och den energimängd vattnet introducerar i firnpacken, särskilt vid återfrysning av vattnet, och avgivande av latent värme. Detta ger att temperaturförändringarna kan användas för att studera mönstret av infiltration av vatten genom firnpacken. Data från de tre sommarsäsongerna visar:

- en hög lateral inhomogenitet av vatteninfiltration;
- en hög uppvärmning av varje smältsäsong i firnpacken pga av avgivande av latent värme från återfrysande smältvatten;
- en hög uppvärmningstakt av firnpacken under mätserien där de övre 10 till 12 m av firnpacken värmdes upp under observationsperioden på tre år från säsongsvis kall till av vara helt tempererad.

De två senare resultaten kunde inte simuleras genom den algoritm som upprättades för att beräkna värmeflödena i firnen med hjälp av en standardkonfiguration för beräkningar av perkolerande vatten i en firnpacke.

Motiverad av resultaten från studien av preferentiell dränering implementerades en rutin som beskriver processen i firnpackealgoritmen. Det föreslagna tillvägagångssättet är att först fördela vattnet som bildas på ytan vid varje tidssteg till de olika underliggande firnlagren. Därefter exekveras den konventionella vattenperkoleringsrutinen. Vattendistributionen längs profilen styrs av perkolationsfunktionerna som minskar vattenflödet med djupet.

Resultat föreslår att det för varje perkolationsmönster finns ett optimalt perkoleringsdjup. Detta bekräftas av temperatursimulationen genom en bättre överensstämmelse med observationerna när denna modell används. Osäkerheterna med denna typ av beräkningar är ca två gånger lägre jämfört med resultaten från standardschemat. Mekanismen för den föreslagna modellen av det preferentiella vattenflödet är baserat på en underskattning av potentialen hos kapillära krafterna att hålla kvar vatten i balans mot gravitationen. Det visades att med en rimlig ökning av smältning vid ytan förändrades inte korrelationen mellan simulerade och observerade energinivåer i firnpacken, vilket ökar trovärdigheten av den modell som användes i detta arbete.

Slutsatsen från detta delarbete är att det preferentiella vattenflödet i firn kan simuleras med den endimensionella flerskiktsmodell som presenteras i avhandlingen med ett modifierat schema för hur vatten perkolerar genom olika lager. För att ytterligare förbättra modellen krävs dock en ny serie fältobservationer.

Den del av data som registrerades under de kalla årstiderna användes för att beräkna optimerade värden av effektiv värmeledningsförmåga (k_{eff}) i firnprofilen. Tre temperaturserier där temperaturen inte överstiger -2°C (våren 2014, hösten – vinter 2014, våren 2015) skapade uppfyllde gränsvillkoret och en värmeledningsmodell baserad på Fouriers lag kunde skapas. I arbetet skapades tre modelldomänerna på ca 2 m tjocka skikt ur vilka det erhölls ett värde på k_{eff} ur värmeledningsekvationen. De k_{eff} värdena användes sedan för att minimera felet mellan de simulerade och uppmätta underytemperaturerna med hjälp av en optimeringsrutin. Ytterligare numeriska experiment gjordes för att undersöka de potentiella osäkerheterna i uppskattningarna av de optimerade k_{eff} värdena som kan vara orsakade av fel i det empiriska data som används.

De optimerade k_{eff} värdena ökar med djup och varierar från 0.6 till 1.3 $J\ (s\ m\ K)^{-1}$. Dessa värden är konsekventa och signifikant högre än de resultat från de allmänt använda densitetsbaserade parametriseringar som föreslagits av Sturm et al. (1997) och Calonne et al. (2011). Detta tyder på en underskattning av konduktion och effektiviteten av värmeledningen av de densitetsbaserade parametriseringar som tidigare föreslagits.

De optimerade k_{eff} värden som togs fram för firnlagret på Lomonosovfonna användes sedan för att beräkna mängden porvatten i firn efter ablationsäsongen 2014, baserat på utvecklingen av fintemperaturer mellan september 2014–januari 2015. För att möjliggöra denna uträkning föreslås en ny metod.

Hypotesen för denna delstudie är att firntemperaturens utveckling under den kalla delen av året kan definieras som två ledande processer: 1) återfrysning av det vatten som hålls kvar i porerna efter ablationssäsongen och, 2) konduktion av värmeflödet mot köldkällan. Som ett resultat av dessa processer tränger frysfronten nedåt i firnpacken med tiden. Detta kan kvantifieras från den temperaturutvecklingen i firnpacken. Med detta föreslås det att vattenhalten i firnen kan beräknas med hjälp av den uppskattade effektiviteten i konduktionen och av frysfrontens vertikala förändring i firnpacken.

Den resulterande porvattenhalten i denna studie är ca 1–2.5 $vol.\%$ för profilens övre del (0–3 m djup refererad till glaciärytan i april 2015) och mindre än 0.5 $vol.\%$ under 3 m djup. Dessa värden är signifikant lägre än de resultat som fås ur den parametrisering som är baserad på firndensitet (Schneider and Jansson, 2004). Detta resultat och det faktum att det var stora laterala variationer i temperaturprofilerna tolkades som konsekvens av fickor av lägre porvattenvolymer i detta arbete som ett bevis för preferentiellt vattenflöde.

Fältobservationerna och de kvantitativa resultaten i avhandlingen ger ytterligare empiriska bevis för preferentiellt vattenflödet i snö- och firnlager. De resultat som presenteras i avhandlingen visar att det krävs att skiktade firnmodeller kunna beskriva den där processen för att beräkna flödet av massa och energi noga. Detta kommer att resultera i en bättre förståelse för hur avrinning av smältvatten ur glaciärer och inlandsisar påverkas av pågående och framtida klimatförändringar.

8. Acknowledgments

The author of the present thesis is convinced that the this chapter is exceed in length by the other sections for the sole reason of its non-scientific nature. By no means does the fact that he is the lead author on the four papers, forming the core of the present thesis, diminish the role of numerous individuals and organizations in the work on both the papers and the thesis as a whole.

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As it was stated in several places above, the thesis is to a large extent based on data collected during six field campaigns (2012–2017) at Lomonosovfonna in Svalbard and consecutive work in the UNIS and NPI cold laboratories. The field site, although placed in a rather flat and environmentally challenging land(ice)scape, was never a boring place thanks to the great people inhabiting this snowy spot time to time. I appreciate the cheerful mood and constructive help of Veijo, Rickard and Ward mentioned above, but also of Horst Machguth, Christian Zdanowicz, Dorothée Vallot, Wim Boot, Katrin Lindbäck, Carmen Vega, William Kohler, Glennda Villanflor and Elisabeth Isaksson. It was always impressive that despite the multiple "surprises" like snow scooters turning upside down and leaking oil, 10 *Ohm* resistors in place of 10 *kOhm*, scooter sledges falling apart, flat batteries and firn cores shipped to wrong labs, somehow, things worked out in the end. I have to say that a lot of practical knowledge was accumulated during these time- and resources-wise efficient and logistically careful field campaigns and will, hopefully, have the chance to apply and share that experience.

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