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Report

Review of internal accumulation rates on glaciers of Svalbard

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Contents

Abstract

Climatic and glaciological conditions at glaciers of Spitsbergen favor developments of processes of internal accumulation. It affects glacier mass and energy balance and also their internal structure both in terms of composition and temperature. Numerous studies have been made on glaciers of Spitsbergen to estimate the qualitative and quantitative characteristics of internal accumulation. In the current report a review of published results on rates and type of internal accumulation at several glaciers of Spitsbergen are presented. The report covers following glaciers and ice caps Midtre Lovénbreen, Storöyjökulen, Kongsvegen, Vestfonna and Nordenskiöldbreen. Typical values of refreezing range from 0.15 to 0.35 mwe. with substantial spatial and temporal variability. Driven by apparent difficulties in direct field measurements of refreezing rates a number of scientists have applied modelling approach to resolve for that component of the mass balance. We undertake an attempt to estimate the performance of a layered physically based englacial model (SOMARS) by testing its output against a field dataset from Lomonosovfonna. The model reproduces englacial temperature and density profile reasonable well in most part of the accumulation zone, but largely overestimates subsurface densities. We link this inconsistency with a possible error in simulated refreezing rates of 20-27%.

Introduction

The landscape of the Svalbard archipelago (Fig. 1) is dominated by glaciers and ice caps, with a cover of ca. 36600 km^2 , which makes up for approximately 7000 km^3 and 60% of the area of the archipelago (*Hagen et al, 2003a*). This extent of glaciation has a large effect on all systems on Svalbard, affecting the climate on regional and local scales, shaping the landforms and producing sediments, runoff from glaciers serves as the primary source of water for the surface water systems on the islands and also influences the stratification of water in local fjords. Ultimately the water drained from the glaciers on Spitsbergen contributes mass to the world ocean system and in a longer perspective influences the global sea level.

Being located in a high latitude zone of the Arctic Spitsbergen archipelago is characterized by a harsh climate with cold winters and modest rates of precipitation. Summer season is normally very short, but due to the influence of warm North Atlantic current it still longer than in other regions lying at high latitude 78-80°N. In the recent study presented by Førland et al (*2011*) the mean annual air temperatures for Longyearbyen airport weather station (central Spitsbergen) are estimated at -6.7°C and -4.6°C for standard normal periods 1961-1990 and 1981-2010 correspondingly, which results in an increase of 2.1°C, most of which is attributed to the winter and spring seasons. Estimations of the precipitation rates close to the sea level are more uncertain and spatially variable: mean annual precipitation at Longyearbyen airport and Ny-Ålesund for standard normal period 1961-1990 is 190 mm and 385 mm correspondingly while in 1981-2010 the values are 191 and 427 mm (*Førland et al., 2011*).

A number of glacier mass balance estimates and climatic assessments from Spitsbergen are available from conventional measurements and ice core studies (*Hagen et al, 1990; 2003a; 2003b, Zemp et al, 2011, Divine et al, 2011*, see Fig. 2). The general pattern in evolution of mean specific surface mass balance of glaciers of the archipelago was found to be negative in the second half of the 20th century: -14±3 mm/yr, which if also accounting for iceberg calving corresponds to the rate of global sea level rise

0.01 mm/yr (*Hagen et al, 2003b*). It is important to note that conventional measurements of ablation and accumulation on glaciers and ice core analysis are available from only few glaciers of the archipelago. In the perspective results drawn from remote sensing data become particularly relevant, since they allow for a great spatial extent.

According to recent estimates (*Nuth et al., 2010; Moholdt et al, 2010a, 2010b*) of spatial and temporal dynamics in glacier surface elevations at Spitsbergen during the last 15-40 years the dominant pattern was thinning at lower altitudes, while at high elevations glaciers were thickening or stagnant. The geodetic mass balance (excluding calving) of Svalbard glaciers (excluding Austfonna and Kvitøa) during 1965-2007 yy. was suggested to be -360±20 mm w. e./yr which corresponds to 0.026 mm/yr of global sea level rise (*Nuth et al, 2010*). Later analysis of laser altimetry data showed that glaciers of Svalbard have been loosing on average 120±40 mm w. e./year during 2003 - 2008 (*Moholdt et al., 2010a*). According to one of the most recent estimates made on the basis of conventional glaciological measurements, satellite gravimetry and laser altimetry glaciers of Svalbard have been loosing 130±60 kg/m^{2}y during 2003-2009 yy. (*Gardner et al., 2013*).

Figure 1. Locations of glaciers from which published data on internal accumulation rates is discussed further in the report.

It is also worth noting that there is a number of uncertainties in estimation of rates of glacier ice loss and gain. Often correlations between results derived by different methods is poor. One of the major limitations is the lack of data on density-depth distribution and its change over time. In both cases: whether mass balance estimates are drawn from conventional mass balance measurements or from comparing data on glacier surface elevation for different moments in time, change in mass below the summer surface in the former case and glacier surface in the latter has to be assumed from a prescribed density distribution (*Hagen et al., 2003a; 2003b*). As a consequence mass balance estimates for the accumulation zone, where subsurface density can experience changes, are less reliable. In that perspective mass balance records reconstructed from ice cores taken above the runoff

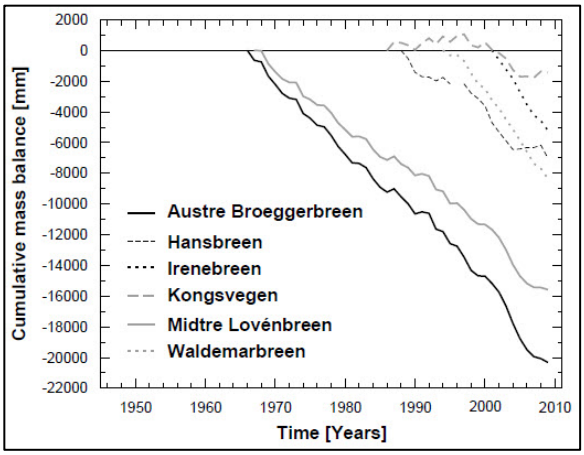

Figure 2. Cumulative glacier mass balance series from direct measurements at Spitsbergen (*Zemp et al, 2011*).

limit are a better estimate of glacier evolution as when averaged over several years this data is a direct source of information on mass accumulated during a given period.

As it was noted earlier, remote sensing data is particularly important for studying glaciology of Svalbard glaciers. One parameter commonly used to estimate the mass balance of glaciers is the altitude of the equilibrium line - ELA (*Cuffey and Paterson, 2010*). However, in the Arctic and sub-Arctic regimes position of the equilibrium line is offset from the easily identifiable snow line by the superimposed ice zone (Fig. 3). Thus formation of superimposed ice is a factor that introduces an appreciable uncertainty in mass balance estimates drawn from temporal dynamics of ELA (*Engeset et al, 2002*).

Thus majority of mass balance estimates are to a large extent affected by the processes involved in internal accumulation and formation of superimposed ice, which is recognized as an important component of the glacier mass balance at Spitsbergen (*Hagen et al., 2003a; 2003b*). In extreme cases glaciers can receive up to 100% of annual accumulation in the form of superimposed ice (*Jonsson and Hansson, 1990*). This may be explained primarily by very low annual and winter temperatures, which increases the cold content of the active layer by the end of accumulation season, and relatively low rates of accumulation which decreases the insulating effect of the snow layer, reduces pore space and leads to higher relative importance of refreezing.

The purpose of present report is to give a qualitative description of processes involved in internal accumulation on glaciers of Spitsbergen and environmental parameters controlling it. We would also like to review and discuss published data on rates of internal accumulation at glaciers of Spitsbergen. It is important to understand what magnitude with respect to other components of mass and energy balance the process has. Driven by the fact that precise field measurement of melt water refreezing rates in the field is a challenging task significant efforts were also undertaken to model the process on the basis of other environmental variables. In the current report we also undertake an attempt to assess the performance of an englacial model widely used for description of evolution of density and temperature in glaciers and applied at Lomonosovfonna at Spitsbergen (*van Pelt et al., 2012*). Locations of glaciers discussed further in the text are shown in Fig. 1.

Theoretical background

According to the Glossary of Glacier Mass Balance and Related Terms (*Cogley et al., 2011,* Fig. 1) ”refreezing of water within a glacier, between the summer surface and the bed, which goes undetected by measurements of surface mass balance” is called ”internal accumulation”. It would be logical to consider the problem in the framework of a larger section of glaciology – mass and energy balance of glaciers.

Internal accumulation may be realized by different processes. It can occur as refreezing of water in the pores of snow and firn either in spring when warming of the subsurface layers occurs, or in autumn, when water suspended in pores refreezes as the cold winter wave penetrates down from the surface. Another option is that melt-water refreezes directly on top of the solid ice and forms a layer of superimposed ice (*Reijmer et al., 2012*).

Components of glacier surface mass and energy balance, and internal accumulation in particular, are controlled by the processes in the lower troposphere. This problem from the most general positions was thoroughly reviewed by Shumskii (*1955, 1964*), Benson (*1960, 1961*), and Muller (*1962*), as a result the study of ice formation zones or glacier facies was developed (Fig. 3). Freezing of liquid water can

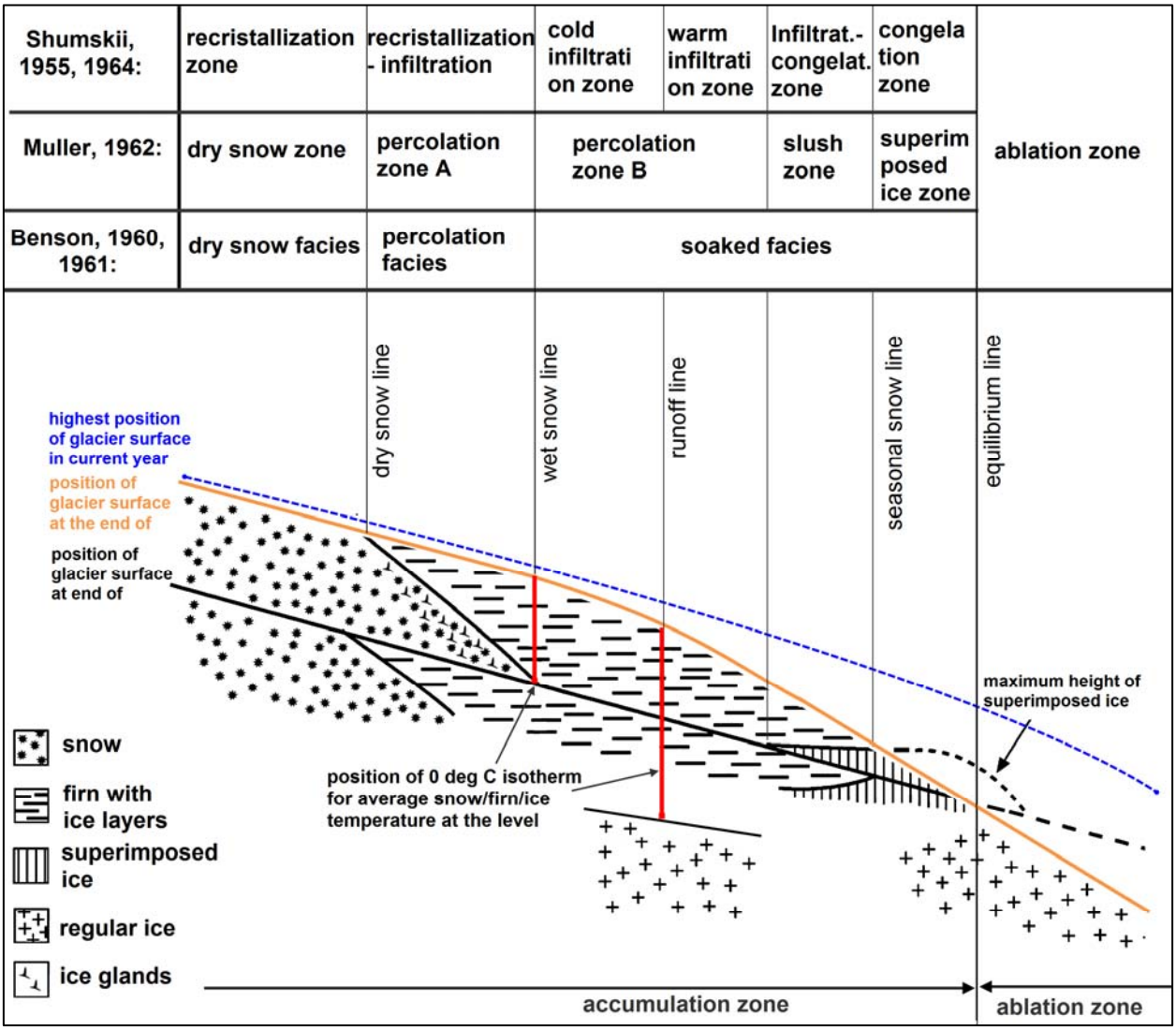

Figure 3. Zones of a glacier based on Shumskii *(1955, 1964)*, Benson *(1960, 1961)* and Müller *(1962).*

occur in all ice formation zones, but the particular processes involved and their magnitude are controlled by the balance of mass and heat at the glacier surface. Details on characteristics of each glacier zone were discussed in our earlier report ("Internal accumulation on glaciers: qualitative description and quantitative estimates") and here we will shortly touch upon details particularly relevant for glaciers of Spitsbergen.

Owing to the vast variety of climatic and glaciological conditions on Spitsbergen and to the fact that many glaciers lie in a large range of altitudes, glaciers of the archipelago appear in different zones and thus experience different processes of internal accumulation. Many glaciers at Spitsbergen are polythermal and are characterized by a layered internal thermal structure, which can be explained by active melt water refreezing and ice dynamics. A sketch qualitatively presenting general spatial patter of distribution of glacier zones in the northern part of the Spitsbergen island is shown in Fig. 4 (*Zagorodnov, 1985*). One can note that the figure apparently does not show any zones of temperate ice, which is its shortcoming. Below we present an overview of glacier zones at Svalbard with reference to Fig. 3 and to particular glaciers further discussed in the report.

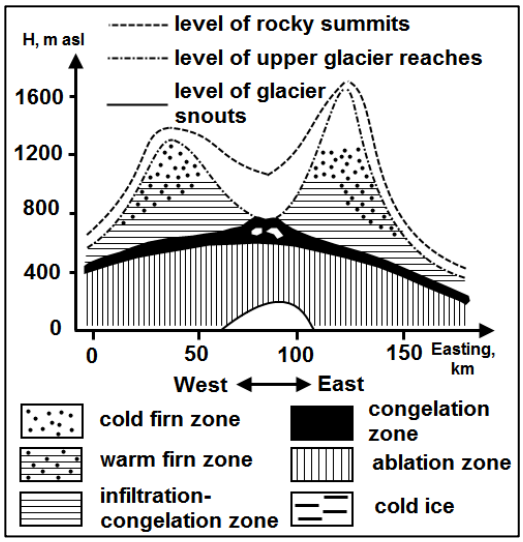

Figure 4. Scheme of glaciological zonation at glaciers of Spitsbergen along the profile at 79 °N (see Fig. 1 for location of the profile) *(Zagorodnov, 1985).*

Under cold and relatively precipitation rich conditions the availability of melt water is low and it entirely refreezes inside the upper porous layers of snow and firn without reaching the depth where annual amplitudes are negligible (10-12 m). These areas belong to the *cold infiltration* zone according to the classification by Shumskii (1955, 1964) and found on the East of the archipelago at elevations higher than 900-1200 m and at Vestfonna and Austfonna ice caps at somewhat lower belt (above 550 m) (*Zagorodnov, 1985; Dunse et al., 2009*). Melt water here affects the density and temperature of snow and firn increasing both and is important to be accounted for. Refreezing happens both during the spring and autumn seasons. Process of water refreezing and release of the latent heat is especially important under conditions of warming climate when increasing melt rates may not result in increased run off because the energy will be spent on warming the glacier up by refreezing. But once the cold content is used the part of the glacier will exit the domain of cold infiltration zone and will start to contribute to the runoff. According to the results of temperature measurements (*Van de Wal et al., 2003*) and stratigraphical studies (*Pohjola et al., 2002*) upper reaches of Lomonosovfonna are also in the cold infiltration zone. It may be noted that an inverse temperature gradient is observed in the interval 15-70 m, which implies a substantial warming during the last 100 years with the rate of 0.02 – 0.025 K/y (*Van de Wal et al., 2003*). In case this trend is maintained it is likely that the area will shift to the warm infiltration zone.

At lower elevations where warmer climatic conditions are observed amount of melt water increases and it is also probable that precipitation will partly consist of liquid water. Runoff occurs when either the cold content or the pose space become a restriction for refreezing of all available liquid water.

In the first case the upper part of the snow-firn-ice column which is a subject for seasonal changes of temperature is not cooled enough during winter season to ensure refreezing of all the liquid water

available. Melt water percolates deeper than the depth of zero year amplitudes and warms the snow-firn column up to the melting point. After that the snow/firn pack can still hold some additional water and prevent it from running of by capillarity forces. But if melting continues runoff is initiated and water evacuates through englacial and subglacial conduits and channels. The glacier in this case is a warm glacier (or its part in question is warm). Similar conditions are observed on a large number of glaciers at Spitsbergen, especially in the southern and western parts of the archipelago, characteristic range of latitudes is 450-850 m. Warm infiltration conditions were described at Isachsenfonna, Amundsenisen, Werenskioldbreen at the ice divide of Grønfjordbreane and Fritjovbreen (*Zagarodnov, 1985*), at in the accumulation zone of Kongsvegen, Uvérsbreen, Midtre Lovénbreen (*Björnsson et al, 1996*), Hansbreen (*Jania et al., 1996*).

In the second case the relation between sum of melt and rain water and amount of solid precipitation is such that refreezing water fills entirely the pores and a slush layer accumulates on top of the impermeable horizon. Assuming that the cold content of the active layer is large, refreezing will occur at the ice-slush contact and a layer of superimposed ice will form. On top of that newly formed layer runoff will occur. Conditions similar to those described above are indicators of infiltration-congelation zone and on Svalbard are generally observed on glaciers in the altitude range 380-600 m (*Zagorodnov, 1985*). Reliable data has been reported from Holtedahlfonna (*Troitskiy et al., 1975*) and Lomonosovfonna (*van Pelt et al., 2012*).

Under conditions of warmer climate and smaller precipitation rates, the layer of firn left on the glacier by the end of the accumulation season decreases. In the lowermost part of the accumulation zone not only the seasonal snow but also a part of newly formed layer of superimposed ice may be ablated. This happens in the *congelation zone* where the main process involved in internal accumulation is formation of superimposed ice during summer. The main requirements are that the surface energy balance in the winter season is considerably negative to ensure a steep temperature gradient in the ice and that the accumulation rates and consequently the pore space available are relatively low. In extreme case the entire glacier can be nourished by superimposed ice, which is the case at Storöyjökulen in the North-East of the archipelago (*Jonsson, 1982; Jonsson and Hansson, 1990*). But in most cases the extent of congelation or superimposed-ice zone is much smaller – it stretches vertically 50-200 m above the equilibrium line of glaciers.

It is to be noted that internal accumulation is significant for mass and energy balance of all parts of a glacier. Even though in the ablation zone the role of refreezing is not as apparent as in the accumulation zone, it indirectly affects the local energy and mass budget. The latter two are closely linked and dependent on each other. Internal accumulation affects not only the relation between evident lowering of the glacier surface and the amount of runoff, there is also an indirect effect. it is obvious that if by the time melting starts the snow/firn/ice pack is temperate, then the entire heat flux from the atmosphere would be spent on lowering the glacier surface by melt. In most cases this is not the case and ablation is reduced by refreezing. Thus even in *ablation zone* it is important to account for refreezing of liquid water because without that the net effect of melt season may be overestimated.

Reported rates of refreezing at glaciers of Svalbard

Midtre Lovénbreen

Midtre Lovénbreen is a small valley glacier (area around 5 km^2) in the North-Western Spitsbergen (Fig. 5). This glacier has a well established series of mass balance observations according to which the average net mass balance in the period 1968-2002 was -0.33 m w. e./year, which corresponds to -12.11 m w. e. of cumulative total mass balance. The terminus is retreating since the first half of the 19th century (*Wright, 2005*).

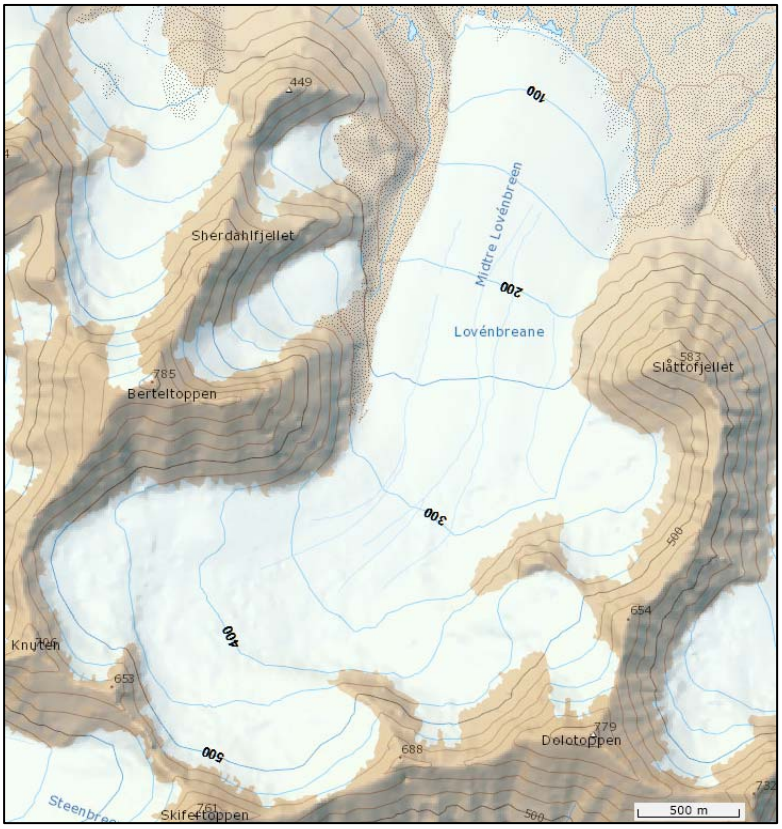

Figure 5. Map of Midtre Lovenbreen (NPI)

According to the field based results from the glacier its mass balance is highly dependent on internal accumulation (*Wadham and Nuttall, 2002*). From September 1998 to September 1999 air temperature, snow height and temperatures were logged in two locations on the glacier (160, 395 m a. s. l.). During the summer season density measurement and stratigraphical descriptions of snowpack in pits were done at elevations 160, 260 and 395 m asl. Warm spells, observed on both stations in November 1998, resulted in melt followed by refreezing in pores and on top of the summer surface, which was estimated at 0.18, 0,17 and 0,11 m w. e. at altitudes of 160, 260 and 395 m a s. l. correspondingly. Taking into consideration the typical accumulation rates at the sites these values make 40% of winter accumulation for the lower location and 6% of the glacier wide winter balance in 1998/99 season. The layer of superimposed ice formed during the following melt season of 1999 was found to be increasing with elevations and made 16% of the winter accumulation at Midre Lovénbreen (*Wadham and Nuttall, 2002*).

Andrew Wright in the framework of his dissertation (*2005*) performed a field and modeling experiment specifically aimed at estimating the rate of superimposed ice growth on Midre Lovénbreen.

During the field season of 2002 (26 April – 2 July) a thermistor string was installed on the glacier at 305 m a. s. l. It was used for monitoring the temperature changes which can reflect refreezing events inside the snow column. From comparison of the temperature evolution with output of a 1-dimentional heat conduction model the Wright concludes that refreezing of 11 cm w. e. can compensate for the excess heat in the lower part of the snow profile that the model is not able to reproduce. According to the observations in an ice core, drilled at the site of temperature measurements, during the period 21 cm of ice was superimposed on the summer surface formed in the previous year. Wright et al (*2005*) speculate that the discrepancy between the two results can be explained by refreezing of water in the pores of snow rather than as a homogeneous water layer on top of an impermeable horizon, which would increase the expected height of the resulting layer of superimposed ice. Another reason for the apparent discrepancy can be in an unrealistically high value for ice density used for calculations. Indeed according to observations density of newly formed superimposed ice is usually around 800-900 kg/m^3 (*Koerner, 1970; Jonsson, 1982*), while the pure ice density is much higher - 917 kg/m^3.

By the end of the measurement period the mass of material stored above the previous year's summer surface has not changed much since beginning of the melt season. Due to release of latent heat in the process of refreezing the entire snow profile was isothermal by the end of measurement and liquid water was observed on top of the newly formed ice layer. On the basis of these observations the author draws conclusion that refreezing ceased by the 2nd of July. This conclusion could have been more solid if additional data on temperature gradient inside the firn was presented, as the process of superimposed ice formation is essentially driven by the magnitude of this parameter.

Table 1. Field results of the study on rates of superimposed ice formation on Midtre Lovénbreen in spring and summer 2002 (*Wright, 2005*)

Altitude, m asl	Surface slope, °	Core date	Thickness of superimposed ice, cm	Bulk density of superimposed ice, kg/m^3	Bulk density of glacier ice, kg/m^3	Comments
280	4	3 July	27.25	716	670	Sloping summer surface
290	2-3	3 July	25.5	747	699	Pronounced bubble layers
295	3	3 July	21.5	708	712	
305	4	29 June and 1 July	21	894	846	Average of 3 cores at this site
335	7	29 June	22	906	778	

The author also carried out extensive studies on the rate of superimposed ice growth by taking shallow cores and measuring the distance between the level interpreted as the summer surface of the previous year and ice surface in the current year. These results are presented in the Table 1. No significant trend in rates of superimposed ice with respect to altitude is seen in the field data. Instead of an expected increase in the thickness of the superimposed layer with elevation the apparent tendency is reverse. This is explained by the scarcity of field data as the altitude range of available measurements is rather low and stochastic nature of the phenomena, but it is also possible, that higher rate of water supply at lower elevations is an explanation.

For estimation of the refreezing rates at Midtre Lovénbreen on longer time scales the author uses a comprehensive layered englacial model which has an explicit formulation of superimposed ice growth while lacking a description of firnification. Yet it is not specified in the work whether superimposed ice is allowed to form on bare ice surface or not. The model is based on (*Bassford, 2002*; *Bassford et al, 2005*) which in turn refers to (*Greuell and Konzelmann, 1994*) and (*Wakahama et al, 1976*). It is coupled to an energy balance model and is run for the period from 1971 to 2002. Results revealed the critical importance of internal accumulation in the form of superimposed ice for mass balance of the glacier throughout the whole simulation period. The mean average annual rate of superimposed ice growth is estimated to be 0.22 m w. e./year, which when integrated over the whole glacier domain is $3.95 \cdot 10^5$ m^3 w. e./year or 37% of the total accumulation. It is worth noting that no consistent temporal trend has been identified in the results of the model run. Refreezing rates increase with altitude (Fig. 6 a). As expected the magnitude of refreezing within the snowpack increases with altitude (Fig. 6 b), while the maximum rate of superimposed ice growth is observed at 475 m a. s. l. (Fig. 6 c). The latter is explained by lack of melt water at higher altitudes even though the temperature gradient inside the underlying glacier ice is steep.

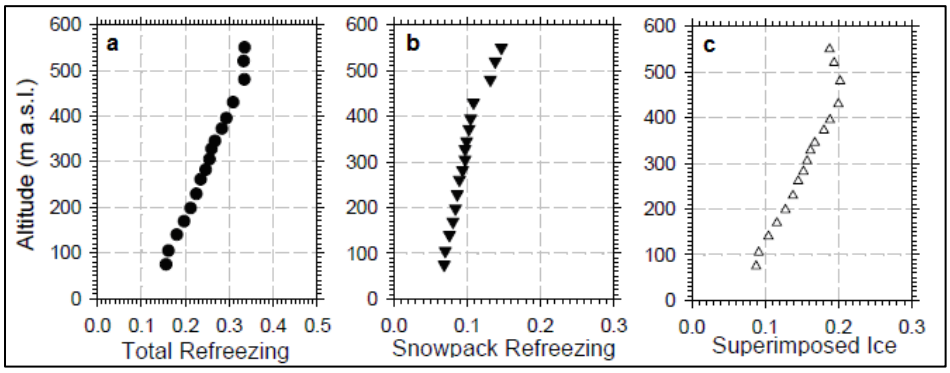

Figure 6. Components of mass balance in [m. w. e./year] plotted against altitude (modified from (*Wright, 2005*).

A common way to estimate or prescribe for modelling purposes the role of internal accumulation for the mass balance of a particular glacier is by introducing the parameter P_{max}, defined as the fraction of precipitation that when expressed in m w. e. sets the limit to melt water refreezing (*Shumskii, 1955, 1964; Reeh, 1991; Pfeffer, 1991; Janssens and Huybrechts, 2000; Reijmer et al.,2012*). The commonly used and cited value of 0.6 for it was offered by (*Reeh, 1991*). In his study A. Wright found significantly lower values for P_{max}: 0.25 - 0.45 at all altitudes on the glacier (Fig. 7), with the maximum reached value of 0.69 and mean value of 0.35. It was also found that the interannual variability of accumulation on the glacier is more pronounced than the variability of superimposed ice formation (Fig. 8). During the simulation period no significant trend in the relation between accumulation and superimposed ice formation which could identify increasing or decreasing importance of the latter was found.

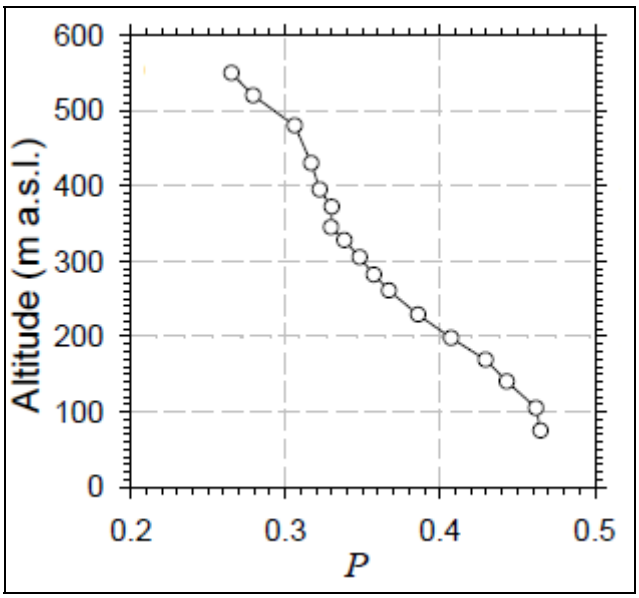

Figure 7. Ratio between accumulation and refreezing (in the snow pores and as superimposed ice) as a function of altitude averaged over the 32 simulation period (Modified from *Wright, 2005*).

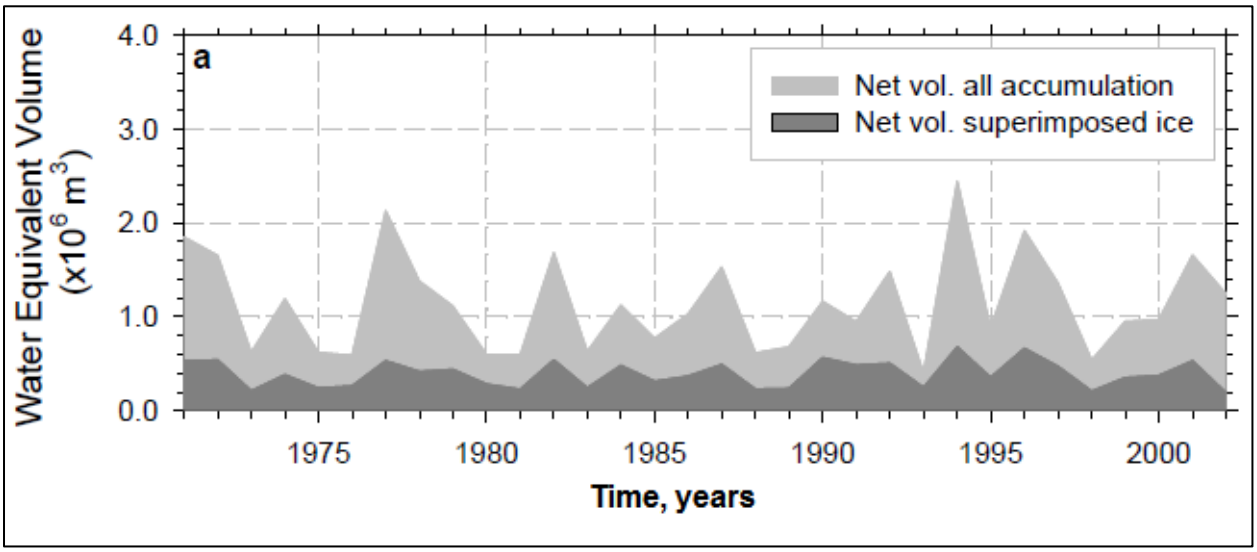

Figure 8. Interannual variability of accumulation and of superimposed ice volume above the equilibrium line on Midre Lovénbreen (from *Wright, 2005*).

In a later study Wright et al (*2005*) compared performance of several models designed for estimation of superimposed ice growth and testing them against the field measurement done on Midre Lovénbreen. The main finding was that any of the six tested models can be tuned within the uncertainty range of the tuning parameter to match the averaged results of observations. Yet three of the models demonstrated a better performance.

Storöyjökulen

In the framework of Ymer-80 expedition in 1980 a small ice cap Storöyjökulen on the North-Eastern Spitsbergen was studied (*Jonsson, 1982; Jonsson and Hansson, 1990*). During the extensive field campaign that was run from July 8 to August 12 a large dataset was collected on the morphology, meteorology and mass balance of the glacier including a unique study on the rates of superimposed ice growth.

Storöyjökulen is a small (ca 50 km^2) ice cap on the Storöya island, lying just east of Nordaustlandet (80.1°N; 28°E). The ice cap covers more than half of the island and is relatively isometric in extent (Fig. 9, left panel). The highest point is 293 m above sea level, to the sides of it the ice cap is sloping gently and reaches the sea level in certain sections of the ice margin – predominantly on the west, south and south east. Storöyjökulen owes its uniqueness to the fact that ice here is formed from snow in one year, meaning that no firn layer is left by the end of the ablation season and the nourishment is realized entirely through the formation of superimposed ice.

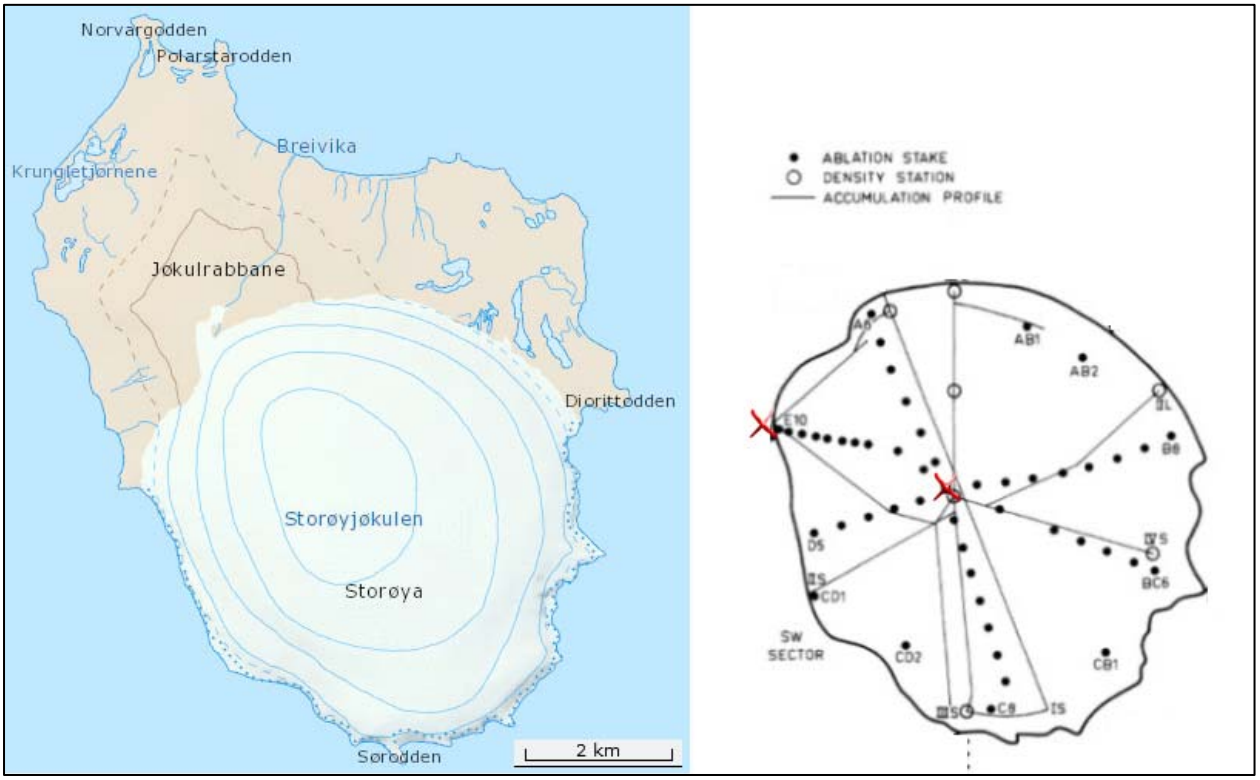

Figure 9. Left panel: *map of Storöya* (*NPI*); right panel: locations of measurement sights, red crosses mark locations of meteorological stations, the one at higher elevation also mark core location.

During the field campaign principal meteorological parameters were observed on two research stations (Fig. 9, right panel): near the western margin of the ice cap (8 m a. s. l.) and on the summit of the ice cap (239 m a. s. l.). For observations on the components of the mass balance a network of 49 stakes stretching in profiles from the top of glacier along the slopes of different aspects was established (Fig. 9). An ice core was derived from the top of ice cap (Fig. 9). From the initial length of 6.8 m only the upper 3.35 m could be reliably used for reconstructions, while the lower part of the core is discontinuous. Comprehensive studies of that ice core ware carried out.

It has to be noted that during the field work neither end of accumulation nor ablation seasons occurred and thus no firm conclusions can be made on the rates of accumulation and ablation on the basis of direct field measurement. Yet the dataset derived provides valuable and unique information on the glaciology of Storöyjökulen. Judging by meteorological data from closest weather stations (Isfjord Radio and Svalbard Airport in Isfjorden and from Hopen island) the year 1980 is representative for the period (*Jonsson and Hansson, 1990*).

Spatial distribution of snow thickness (in m w. e.) measured on the 10-12 July 1980 is presented in the Fig. 10, left panel. Taking into consideration the slow rates of melt in the first weeks of the ablation season authors suggest that it should be very close to net accumulation on the glacier during the 1979/1980 winter season. It can be concluded that the generally low values of annual accumulation (0.1-0.15 m w. e.) are distributed very asymmetrically with higher rates in the upper part of the dome, around the margin and on the south-eastern slope.

The averaged value of ablation for the period July 13 – August 4 is 4 cm w. e., the upper reaches of the ice cap have even experienced accumulation of mass. The spatial pattern of ablation distribution (Fig. 10, central panel) is very asymmetric. The explanation for that is in the method used for calculating ablation and in the large role of summer precipitation. Strictly speaking the central panel of Fig. 6 presents the net surface mass balance of Storöyjökulen during July 13 – August 4, multiplied by -1 to have positive values for mass loss (conclusion one can make from analyzing the formula used for calculation of ablation in original publication). Though on the most of the ice cap (0-150 m a. s. l.) around 0.08-0.1 m w. e. of ice was removed by melt and runoff, during the period close to the top considerable amount of snow was accumulated, which reduced “ablation” both by contributing to the mass and by rising the surface albedo.

The amount of superimposed ice accumulated on top of the summer 1979 surface was estimated from combined measurements of snow depth and distance from the snow surface to the top of each stake in the beginning and end of the observation period (Fig. 10, right panel). Most intensively during the observation period internal accumulation was occurring in the upper part of the ice cap (0.05-0.1 m w. e.) and on its south-eastern slope (0.05 m w. e.). It can be noted that this is also where the accumulation rate is the highest.

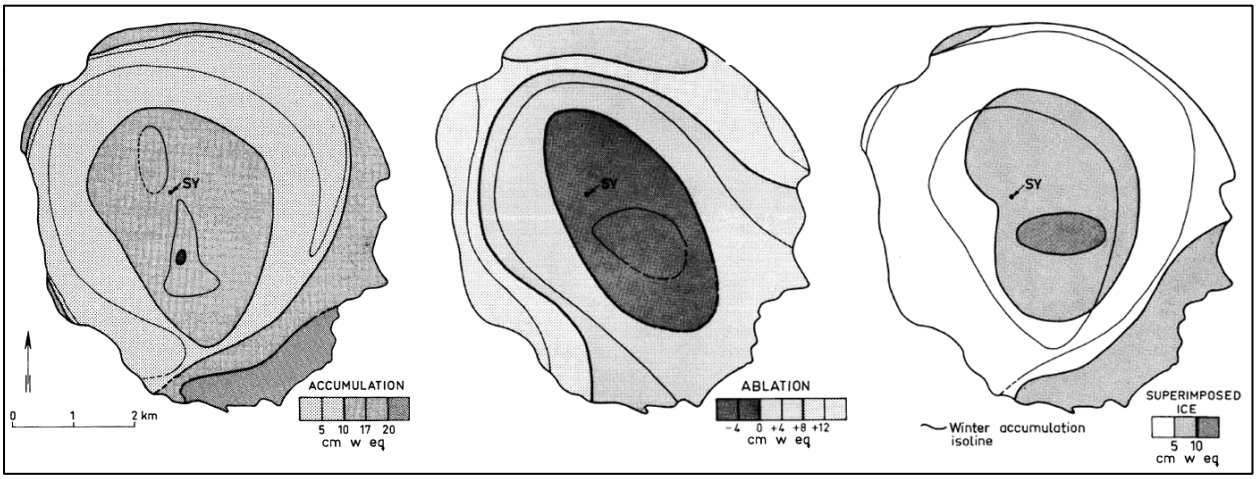

Figure 10. Spatial patterns of accumulation (left panel), ablation (central panel) and superimposed ice formation (right panel) on Storöyjökulen from stake measurements (*Jonsson, 1982*)

Intensive formation of superimposed ice at the coring sight observed during the field campaign allowed to make the assumption that entire annual mass is conserved between summer surfaces of consecutive years and thus annual mass balance can be derived from distance between summer surfaces and ice density. Distance between annual layers in the ice core was used as a proxy for reconstruction of the temporal pattern of mass balance evolution on the top of Storöyjökulen (*Jonsson and Hansson, 1990*). The hypothesis behind splitting the ice core in annual layers was that structural characteristics and composition of ice layers formed early and late in the ablation season should be different as it formation occurs under different conditions. As guiding parameters the authors chose solid electrical conductivity, crystal size and shape, microparticle concentration, occurrence of firn-like structures.

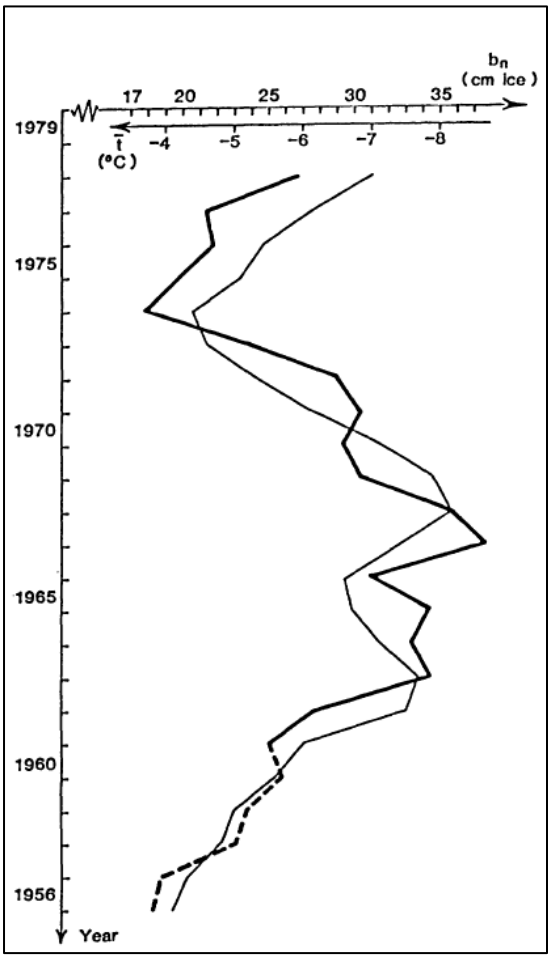

Figure 11. Comparison of thicknesses of annual layers at Storöyjökulen (thick and dashed line) and annual temperatures from Hopen weather station calculated for mass balance years (thin line) (*Jonsson and Hansson, 1990*)

The reliably defined sequence of summer surfaces goes back to 1967 thus providing mass balance estimates back to 1968. Less certain are the estimates for the earlier years of the mass balance history, but still the authors found it possible to go as far back in reconstruction as year 1954 using information from the core pieces available and the regression line between annual temperatures at Hopen station and net balance at the top of the ice cap. The mean value of mass balance in the period 1954-1979 is 0.27 m w. e., much thicker annual layers seem to have been formed in 1963 – 1969 years (Fig. 11).

To test the reconstructed mass balance series against an independent climatic dataset from the same region Jonsson and Hansson (*1990*) compared their results with temperatures at Hopen weather station calculated as mean values for mass balance years (September 1 – August 31). And a prominent correlation was found (Fig. 11). It firstly demonstrates a strong dependence of the rate of superimposed ice growth on air temperatures as the latter are a control on the temperature of the active layer and secondly proves the reliability of the method chosen for determination of mass balance through distance between summer surfaces.

It was further found also that the values of mass balance can be positively correlated with mean oxygen isotopic composition of annual layers (Fig. 12). This correlation, however, was proved to be strong only for the upper part of the ice core (0 – 3.35 m), due to the fact that below that depth ice core record is discontinuous and continuous sampling for oxygen isotopes was not possible. The suggested explanation for the correlation between observed higher $\delta^{18}O$ values in annual layers and years with higher mass balance is that in colder years with higher mass balance values a larger fraction of summer precipitation with heavier isotopic composition refreezes as superimposed ice.

It can be concluded that the study done by Swedish glaciologists at Storöyjökulen ice cap, unique in its nourishment source, presents direct field and laboratory results on the rates of superimposed ice growth, which are very scarce in literature. A combination of very low mean annual temperature, precipitation and melt rates creates conditions for extreme case of internal accumulation on the ice cap.

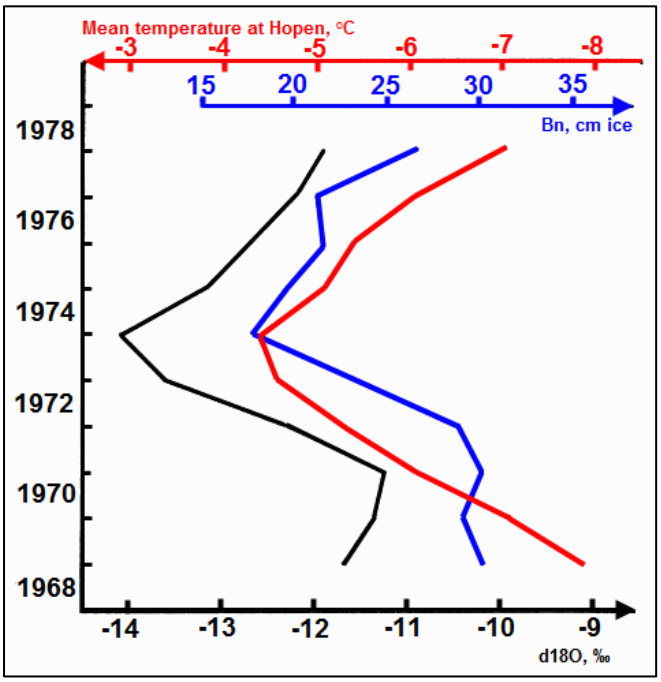

Figure 12. Comparison of thicknesses of annual layers at Storöyjökulen (blue), mean oxygen isotopic signal (black) and annual temperatures from Hopen weather station calculated for mass balance years (red) (*Jonsson and Hansson, 1990)*

Rates of superimposed ice formation has also been reported by (*Dowdeswell and Drewry, 1989*) from the south-eastern slope of Austfonna ice cap that lies close to Storöya. A shallow (9.8 m) ice core was drilled in 1983 at Isdomen (374 m a. s. l.) in the superimposed ice zone. The core was analyzed for tritium concentrations and a horizon with elevated concentrations of the compound was found at the depth of 1 m, from which the authors infer that the rate of superimposed ice growth on the site is 0.047 m w. e./year or 5.1 cm/year.

Kongsvegen

Interesting studies on the rates of superimposed ice formation and its dependence on different meteorological and glaciological variables were done on Kongsvegen glacier, North-Western Spitsbergen (*Obleitner and Lehing, 2004; Karner et al, 2012*). This work was focused on obtaining a dataset on evolution of principle meteorological and englacial parameters from a site on the glacier close to the equilibrium line and replicating the field results by applying comprehensive layered englacial model (*Obleitner and Lehing, 2004).* Further extrapolation of the measured meteorological parameters was done by Karner et al. *(2012*), they used a coupled energy balance and englacial model and presented results on spatial variability of the processes in the boundary layer and below the glacier surface.

Kongsvegen is a large (102 km^2) outlet glacier with a series of mass balance observations from 1986 (Fig. 13). The decade from 1986 to 1997 was marked by an unusual for the region during the second half of the 20th c. positive cumulative mass balance (0.11 m w. e.), while the later decade 2001-2010 exhibited negative values of -1.8 m w. e. (*Karner et al., 2012*). The glacier experienced a surge in 1948, after which is in a quiescent phase.

Figure 13 Map of Kongsvegen glacier, location of the weather station is marked with a red cross

Obleitner and Lehning used a physically based layered model Snowpack, originally designed for avalanche studies in Alpine areas (*Bartelt and Lehning, 2002*). The model was applied for simulation of snow/ice profile at a site close to the equilibrium line of the Kongsvegen glacier, where formation and melting of superimposed ice occurs on yearly basis. The model was forced by AWS data from April 2000

to April 2002 that comprises time-series of air temperature, humidity, incoming and outgoing short- and long-wave radiation fluxes, relative surface height as well as of wind speed and direction. The standard configuration of the Snowpack model was supplemented by a routine describing accumulation of water in the snow matrix (slush layer) when runoff is restricted.

Simulation covered two years with very different evolution of snow/ice profile. During the year 2000 melt was initiated relatively late (18 June) and in took 17 days for the wetting front to reach the summer surface of 1999. Superimposed ice was forming in two stages (A and B in Fig. 14). First refreezing of infiltrating water produced a layer of superimposed ice on top of solid ice at exponentially decreasing rate. Thickness of this layer reached 0.15 m after 40 days. Above the newly produced ice a layer of slush was accumulating. That water layer was entirely frozen very fast after cooling in the late August, which made the second stage of internal accumulation during the season. The total simulated thickness of superimposed ice layer was 0.47 m, which correlates closely with the observed in the vicinity of the site value of 0.6±0.1 m. The next ablation season started earlier (9 June). It took 15 days before the wetting front reached the previous summer surface. The thickness of superimposed ice formed was 0.2 m. More intensive ablation during this year caused exposure of the ice surface already on the 17 August. During the rest of the melt season which lasted until 19 September in 2001 the newly formed ice layer and a large part of the superimposed ice left from last year were ablated.

Principal results of modelling efforts were confirmed by observations and measurements. The overall thickness of superimposed ice appears to be slightly underestimated, which can partly be explained by the local variability at the site (*Obleitner and Lehning, 2004*), and also by the fact that according to observations density of newly formed superimposed ice is usually around 800-900 kg/m^3 (*Koerner, 1970; Jonsson, 1982*), while the model assumes that it is much higher - 917 kg/m^3 (author's comment). The model also reproduced reasonably well, though with slight underestimation, one of the most crucial parameters for superimposed ice formation– temperature close to the ice-snow interface. Snowpack model was also able to reproduce a rather fine stratigraphical feature of the snowpack during summer 2000 – a suspended layer of solid ice on top of slush layer, which was confirmed by observations in corresponding autumn.

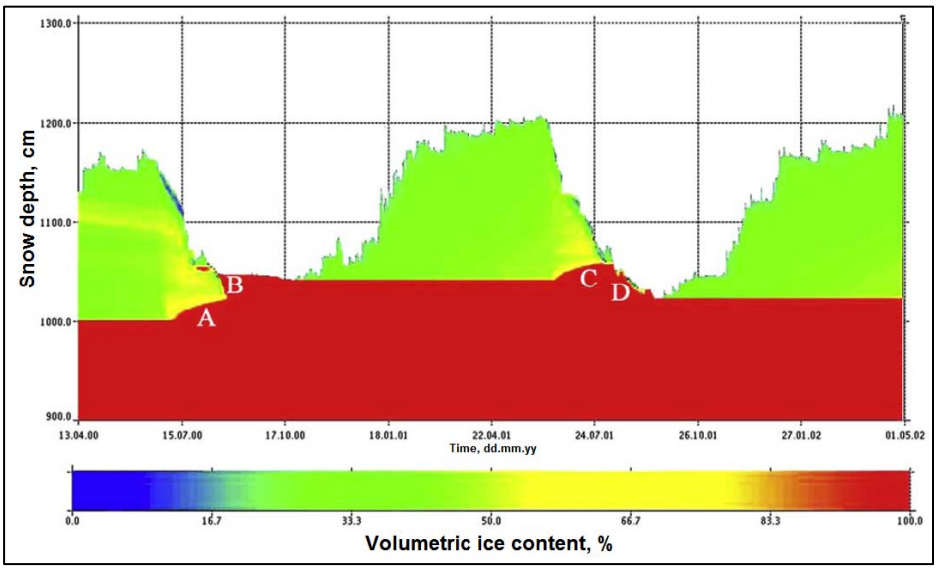

Figure 14. Modeled evolution of volumetric ice content and snow height close to the ELA of Kongsvegen (from *Obleitner and Lehning, 2004*)

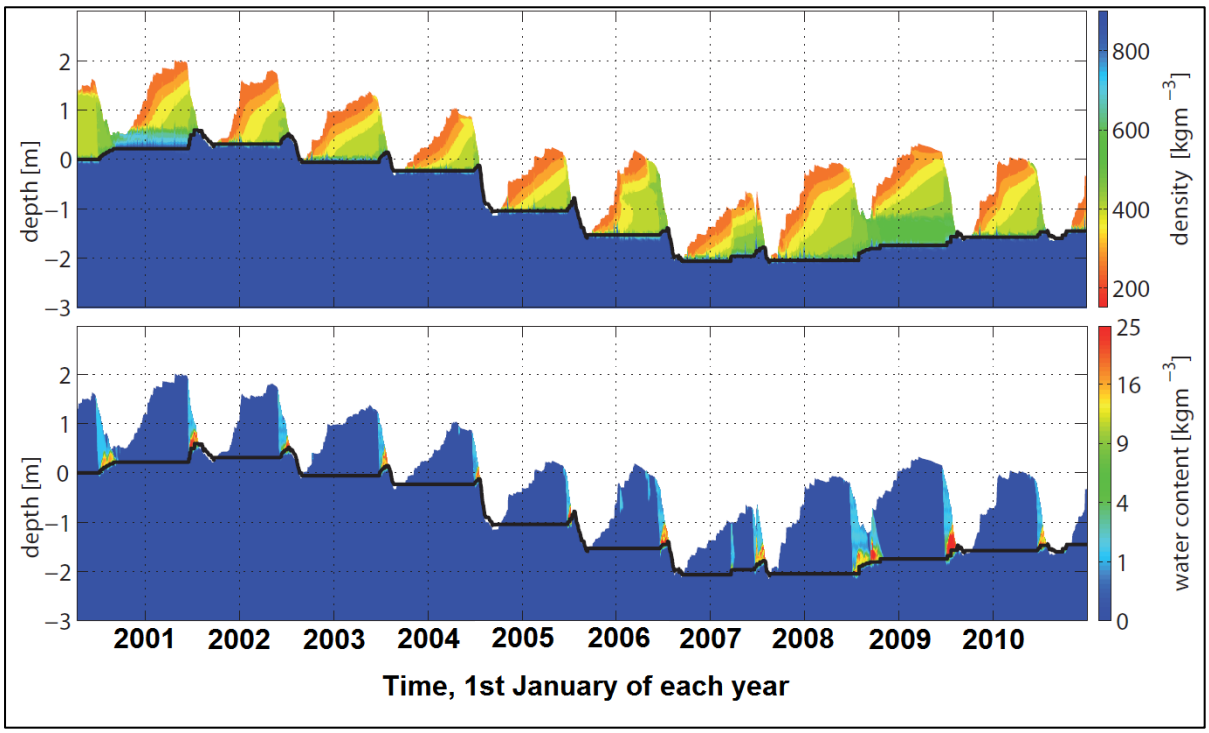

Figure 15. Modeled evolution of englacial density (upper panel) and water content (lower panel) close to the ELA of Kongsvegen (from *Karner et al, 2012*)

These results may be compared with the outcome of a later study done by Karner et al (*2012*) in Fig. 15. This work was based on a dataset from the same weather station on Kongsvegen, but was not focused on internal accumulation, though a thorough description of the corresponding processes was included in the englacial model applied (SOMARS).

It may be noticed that evolution of density profile during summer 2000 and 2001 simulated by two different models is very similar. In both cases the first year (2000) both models produced a layer of superimposed ice during spring melt and a slush layer on top of it. The slush layer was refrozen later during penetration of the winter "cold wave". This might be not so obvious from results of Karner et al (*2012*) as according to their results the modeled density of the refrozen slush layer was less then that in the work by Obleitner and Lehning (*2004*) and has not reached the threshold after which the layer in treated as impermeable. During the second year (2001) the SOMARS simulation used by Karner et al (*2012*) was building the layer of superimposed ice at a much higher rate than the SNOWPACK (*Obleitner and Lehning, 2004*), which can be most probably explained by a larger melt rate produced by the energy balance model in the first part of the ablation season. Yet during the second half of summer the melt rate reported by the later study seem to be smaller and was not sufficient for eliminating not only the ice layer from the previous year, but even the superimposed ice of the current year was not entirely ablated.

Growth of superimposed ice in two stages during early melt season and early winter, which was observed and modeled for the year 2000, was never repeated in the following 10 years. Accretion of ice mainly occurred in the first part of the melt season and resulted in a layer of ice 0.2-0.5 m thick, while in the second part the entire layer was ablated. This, however, was not the case in 2001 and 2007-2010 when the new accumulation season was starting before the accreted layer of superimposed ice was entirely ablated. During these years the study site was in the accumulation zone.

It can be concluded that comprehensive models physically based models, describing evolution of the englacial density and temperature profiles (SNOWPACK and SOMARS) produce reasonable results and

are in generally good agreement with each other. The accurately track principal changes occurring during the melt season but still produce results that are different in details. The largest short coming of these type of efforts is that there is a lack of empirical data against which the simulation results can be tested.

Vestfonna

The history of studies of internal accumulation at Vestfonna is a fine example of international cooperation aimed at scientific progress in describing remote areas and important natural processes that are hard to capture. Major advances were made in the frameworks of two International Polar Years (IPY): during 1956-1958 by expeditions with members from the Soviet Union and Sweden and during 2007-2009 by an international team comprising participants from Germany, Finland, Norway, Poland and Sweden under umbrella of the Kinnvika project.

Vestfonna is a large (area - 2400 km^2) ice cap on the North-East of archipelago with the altitude range from the sea level to 630 m a. s. l. A number of ice streams with calving fronts are draining the ice cap (*Pohjola et al., 2011; Pettersson et al., 2011*). Vestfonna is located to the North-West of another major ice cap of the archipelago – Austfonna, which separates it from the locally major source of moisture in the Barents sea (*Möller et al., 1997*) and lies on the lies on the way of major circulation paths in the lower troposphere (*Beaudon et al., 2011*). Yet recent direct observations on the Vestfonna ice cap have not revealed any statistically significant lateral decrease in accumulation rate from the east to the west (*Möller at al., 2011b*) and according to ice core studies from the upper elevation belts the mass balance of the western summit was almost twice of that on the eastern summit during 1994-2007 (*Beaudon et al., 2011*).

Results of the Third IPY have been reported by Palosuo (*1987*). Then complex investigations on rates of ablation and superimposed ice accumulation were carried out along with detailed studies of glacier subsurface stratigraphy and crystallography. On the western summit of Vestfonna (Ahlmann station) two snow pits were excavated and studied (Fig. 16): in September 1956 (after end of ablation season) and in May 1958 (before start of ablation season), thus the author makes an inference on the rate of ice layer growth during summer 1957. In the earlier pit the summed thickness of ice layers found between 0 and 4 m depth is 17.3 cm, while in the later pit the similar value was measured to be 60.3 cm between depth levels 0 and 6 m. Taking into consideration that the local accumulation derived from the later pit was ca 2.1 m of snow the author concludes that 42.6 cm (0.38 m w. e.) of ice has accumulated during one melt season. The amount of melt, estimated using a simple empirical formula and temperatures at Murchison Bay (western margin of the island, sea level) corrected to account for the elevation difference, was calculated to be 0.37 m w. e. during the summer 1957. From that one can conclude that nearly all meltwater produced at the summit was refrozen during summer 1957.

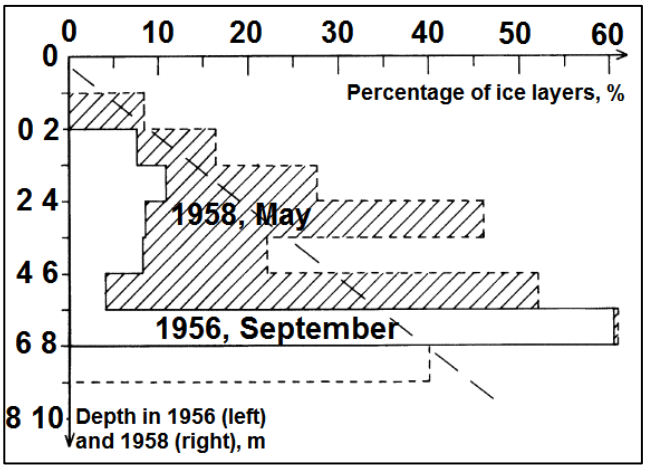

Figure 16. Total thickness of ice layers in each meter derived from snow pits in 1956 and in 1958. (*Palosuo, 1987*)

Palosuo (*1987*) also gives some details on rates of ice superimposition derived by observations and stake measurements on the western slope of Vestfonna during ablation seasons in 1957 and 1958. During the first of these seasons the equilibrium line altitude was estimated as 450 m a. s. l., and the firn line was higher than at least 492 m a. s. l. This means that the altitudinal range of superimposed ice

(congelation) zone was at least 50 that year. During summer 1958 repeated stake measurements gave 2 and 23 cm of superimposed ice accumulation at elevations of 412 and 464 m a. s. l. correspondingly.

Further speculation is made basing on the results of crystallographic study of cores samples. In three cores taken in the elevation range 437 – 478 m a. s. l. an interval with increased size of ice crystals was consistently found at depths 1.5 – 2 m along the profile. Large ice crystals are interpreted as being formed during ponding of melt water on top of impermeable horizons during exceptionally warm summers of 1950 and 1953 (*Palosuo, 1987*).

Results of early studies of englacial temperature, stratigraphy, accumulation and ablation rates (Swedish-Norwegian expedition in 1931 lead by H. Ahlman, expedition of Oxford University in 1935-1936, works during IPY in 1956-1958 described above) were analyzed by Troitskiy et al. (*1975*), which resulted in a scheme of glacier zonation for the accumulation zone of Vestfonna (Fig. 17). It presents a qualitative description of process involved in mass and energy exchange taking place at the ice cap, one of which is refreezing of melt water.

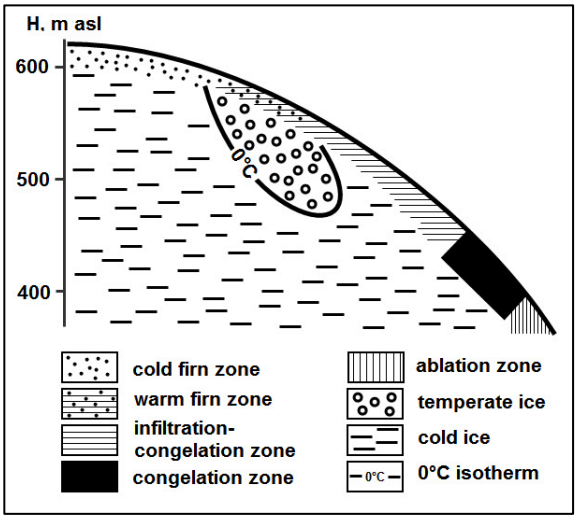

Figure 17. Scheme of glacier zonation at Vestfonna. (*Troitskiy et al, 2011*)

During the Forth IPY three field campaigns were specifically aimed at glaciological studies at the Vestfonna ice cap. They resulted in a number of snow pits studies, 3 ice cores, a dataset on meteorological parameters from a weather station and measurements on a network of stakes. On the basis of this and also other data conclusions were made regarding rates of refreezing rates at Vestfonna.

Taking as input variables ERA-Interim and MODIS datasets Möller et al (*2011a*) applied a set of comprehensive parameterizations to derive spatial patterns of principal mass balance components of Vestfonna during the period from September 2000 to August 2009. The field dataset was used for calibration of the applied parameterizations. This included data series of air temperature and radiation from two AWS, repeated measurements of snow depth and density along with ice surface height changes from a network of 15 stakes and 21 snow pits. For parameterization of refreezing a simple P_{max} method was used (*Reeh, 1991*). According to the approach 60% of the annual accumulation can be refrozen.

According to the results presented by the authors the period-averaged and ice cap integrated values of annual ablation and accumulation are -0.58±0.18 and 40±0.07 m w. e. The refreezing rates were estimated at 0.15±0.04 m w. e./year. On average internal accumulation constituted 25% of ablation rates, ranging in different years from 5 to 48%. Table 2 below presents annual values of precipitation, melt and refreezing for Vestfonna ice cap. From Fig. 18 showing spatial pattern of annual refreezing as a fraction of the ice cap wide mean value it is obvious that the largest rates of refreezing are observed in the upper reaches of the ice cap, where cold content of the firn pack is larger.

Later modelling efforts by Möller et al (*2013*) extended the simulation of climatic mass balance parameters back to 1979 (Fig. 19). Their findings regarding magnitude and temporal evolution of mass balance parameters during 2000-2009 years are in very close agreement with the earlier results. Internal accumulation was found to be one of the principal components of the mass budget are surface mass accounting for 0.21±0.06 m w. e. when integrated over temporal and spatial domain. This corresponds to

refreezing 43% of the annual ablation and 30% of the annual accumulation. The temporal evolution of the simulated value of annually refrozen mass revealed statistically significant trend during the modelling period of 0.026 m w. e./10 y (Fig. 19, lower panel).

Table 2. Annual values ([m w. e.]) of ice cap averaged accumulation, ablation and refreezing at Vestfonna (from *Möller et al, 2011*)

Mass balance year	Annual surface accumulation	Annual surface ablation	Annual Refreezing
2000/2001	0.33±0.05	-0.36±0.15	0.09±0.04
2001/2002	0.29±0.05	-0.55±0.21	0.03±0.03
2002/2003	0.24±0.04	-0.49±0.16	0.09±0.04
2003/2004	0.39±0.07	-0.70±0.18	0.19±0.05
2004/2005	0.51±0.09	-0.78±0.23	0.24±0.06
2005/2006	0.53±0.09	-0.74±0.22	0.23±0.06
2006/2007	0.45±0.07	-0.66±0.20	0.11±0.07
2007/2008	0.51±0.09	-0.45±0.15	0.21±0.06
2008/2009	0.36±0.06	-0.45±0.16	0.13±0.03
Average	0.40±0.07	-0.58±0.18	0.15±0.04
Standard Deviation	0.10	0.15	0.08

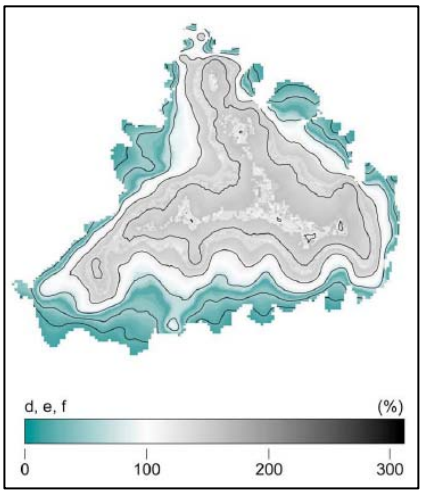

Figure 18. Spatial patter on annual refreezing as a fraction of mean value over the ice cap (from *Möller et al, 2011*)

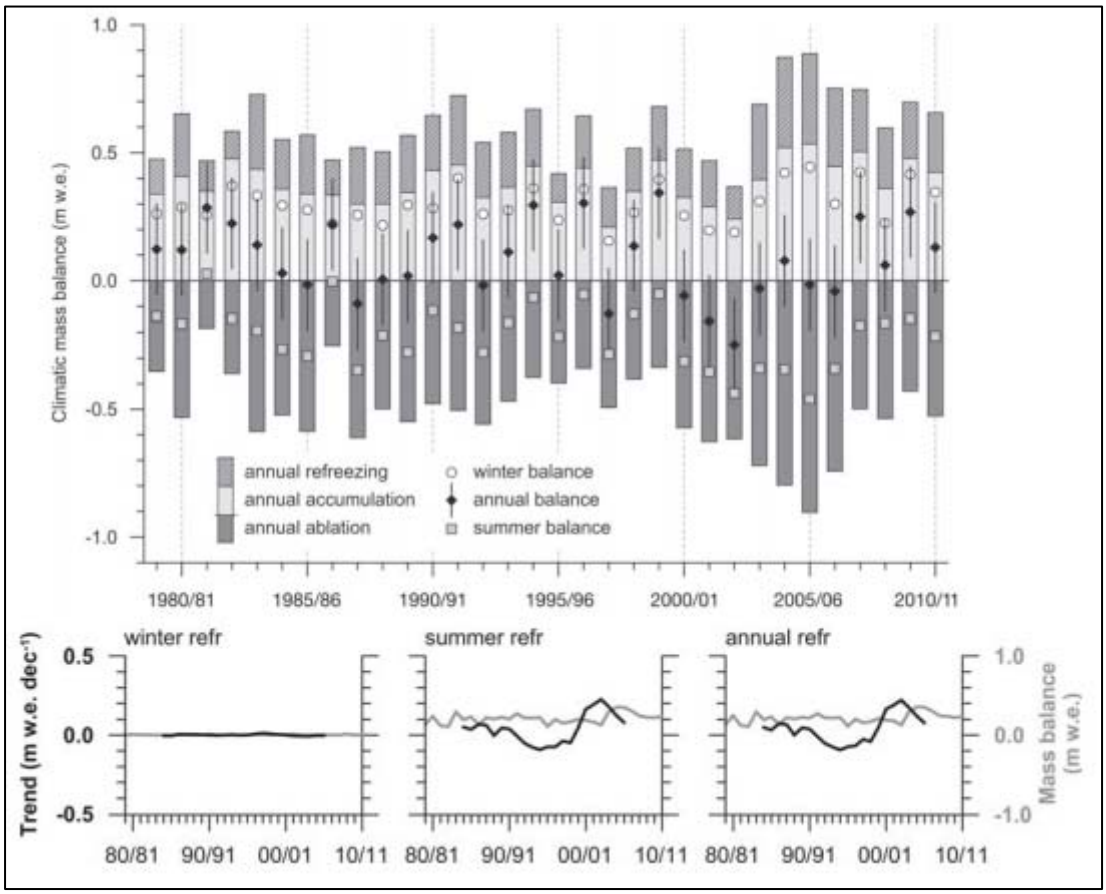

Figure 19. Simulated evolution of main mass balance components (upper panel) and rates of internal accumulation (lower panel) at Vestfonna during 1979-2011.

This last finding goes in line with the evidences from the isotopic composition of ice cores drilled recently on the summits of the Vestfonna ice cap (*Beaudon et al., 2011*). According to Beaudon et al. *(2011*) the snow/firn/ice layers corresponding to the period from 1994 to 2007 had a much heavier isotopic composition than what was reported for the rest of the 20th century. This, together with the significantly positive temporal trend found in the data series itself speaks of a climatic warming occurring at Vestfonna during the 20th and beginning of the 21st century. As a consequence larger melt rates can be expected at the ice cap, which in turn, results in increased refreezing provided that the cold content of the firn layer is sufficient.

From results of radio-echo sounding done at Vestfonna in the framework of Kinnvika project certain conclusions were made regarding the thermal condition in the uppermost part of the ice cap (*Pettersson et al., 2011*). Vast scattering zones were found in the upper 100 m of Vestfonna (Fig. 20), these zones were interpreted of areas of temperate ice. The spatial resolution of early studies at the ice cap in its upper reaches does not allow to make firm conclusions regarding transformation from cold to temperate englacial conditions in the scattering zones that could have taken place since the third IPY, yet there is a high probability that this switch is possible.

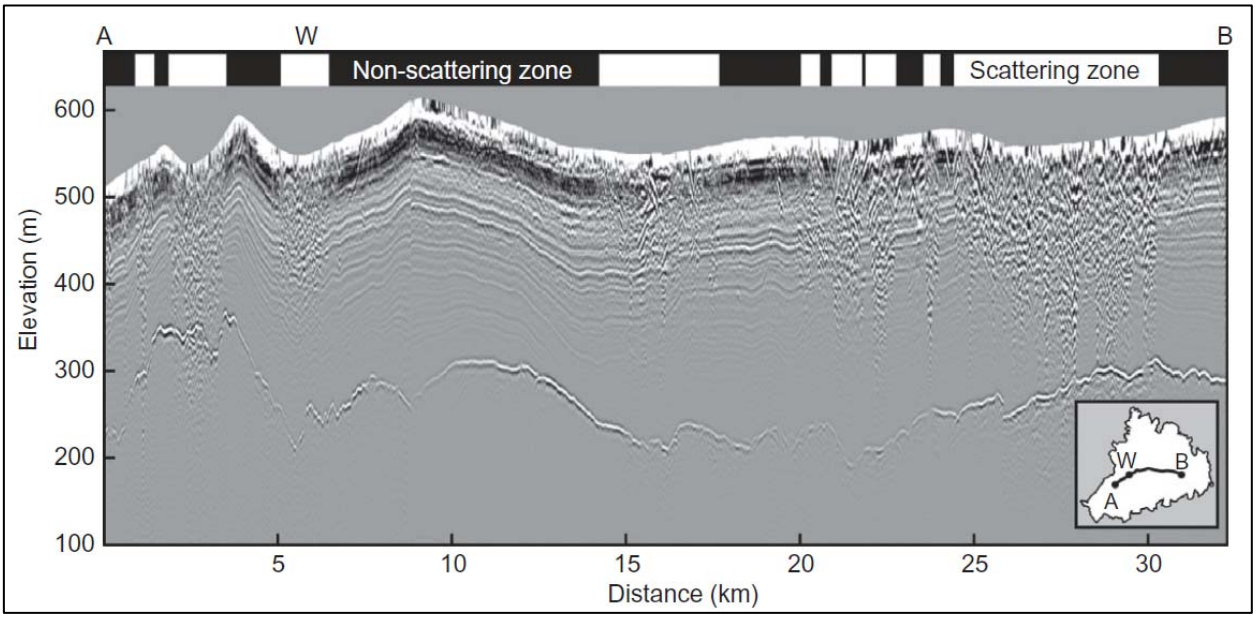

Figure 20 Topographically corrected radar profile along the ice divide of Vestfonna ice cap, location of profile is shown in the lower right corner. The bar in the upper part of the figure indicates localization of scattering zones. (from *Pettersson et al., 2011*)

Thus due to the cold conditions observed at Vestfonna during winter season the mass balance of ice cap largely relies on contribution from internal accumulation. It occurs as accretion of superimposed ice and during refreezing of melt water in the pores of snow and firn.

Nordenskiöldbreen and Lomonosovfonna

Studies aimed at estimation of glacier mass and energy balance at Lomonosovfonna and Nordenskiöldbreen, an outlet glacier thereof (Fig. 21), have been conducted by Soviet glaciologists during summer in 1965. This work was continued later by scientist from Sweden and the Netherlands in the beginning of the 21st century.

During summer 1965 rates of internal accumulation were estimated at 1000 m by two independed methods. From difference between temperatures measured before and after ablation season change of cold content was estimated, it corresponds to refreezing of a 0.1 m layer of water. In the same time, change in cumulative thickness of ice lenses along the profile was estimated at 0.35 m. The conclusion is made that the entire amount of melt water, generated during this summer season at the surface of the glacier, was refrozen in the snow pack.

A comprehensive study of mass and energy balance of Nordenskiöldbreen and Lomonosovfonna was presented by van Pelt et al. (*2012*). The area of glacier domain studied was 193 km^2, spanning a wide range of altitudes: from 0 to 1195 m a. s. l. During the simulation period (1989-2010) the mean net mass balance of the glacier is 0.39 m w. e. with no apparent temporal trend.

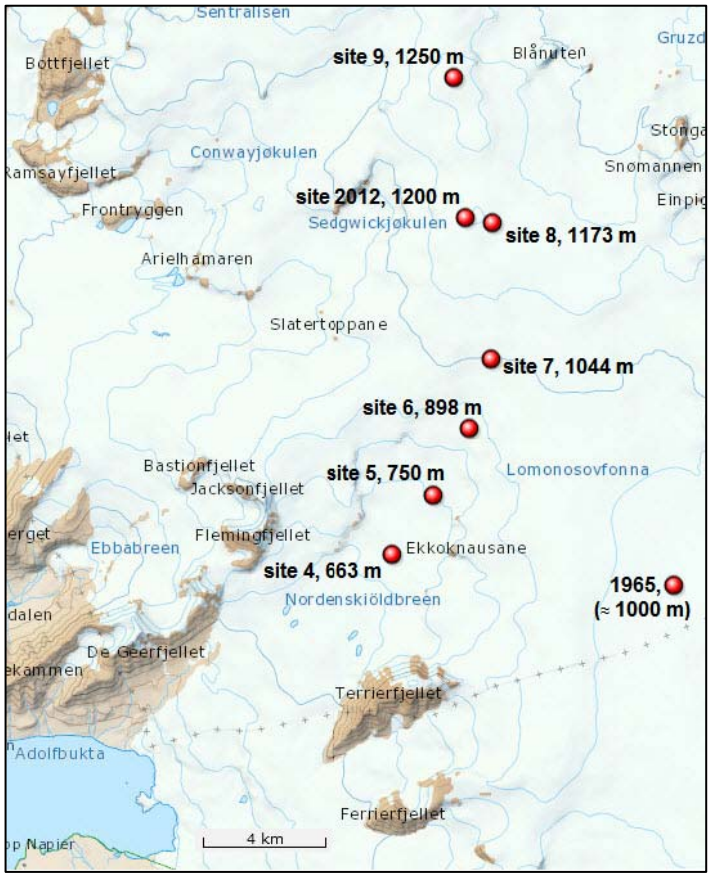

Figure 21. Locations of sites on Nordenskiöldbreen and Lomonosovfonna mentioned in the text. Cores at sites 4, 5, 6, 7, 8. 9 drilled in 1997.

The englacial model (SOMARS) applied by the authors receives input from the surface energy balance model and is coupled to it through the ground heat flux. Though SOMARS is essentially the same model that Wright *(2005)* and Wright et al. *(2005*) used in their studies as both are based on the work by Greuell and Konzelman (*1994*), numerical details of the model applied and its component describing the energy balance are different. Guided by difference in glaciological characteristics between Midre Lovénbreen and Nordenskiöldbreen van Pelt et al. (*2012*) included an accurate description of snow and firn densification by compression and also allow for superimposed ice formation from slush cooled by conduction. The model calculates quantities of temperature, density and water content in 17 horizontal layers with thickness increasing logarithmically and reaching the depth of almost 50 meters. Extensive calibration and initialization of the model is performed before the simulation. It is based on an extensive field dataset obtained at the glacier which includes: radiation measurements by an AWS, snow height observed by an ultrasonic sensor and at 2 stakes, englacial density and temperature profile measured in an ice core.

According to the results of the simulation refreezing is most pronounced in the accumulation zone (Fig. 22), where substantial cold content and large pore space of the active layer define favorable conditions for internal accumulation. It is most intensive at around 1000 m a. s. l., with values reaching >0.35 mwe/year. At higher altitudes the lower melt rates become a limitation, while in the ablation zone ice layer accreted early in the ablation season as melted very soon after snow layer disappears. The rate of refreezing averaged over the spatial and temporal domains is 0.27 mwe/year, which corresponds to 69 % of accumulation rate and just below 25% of the ablation rate.

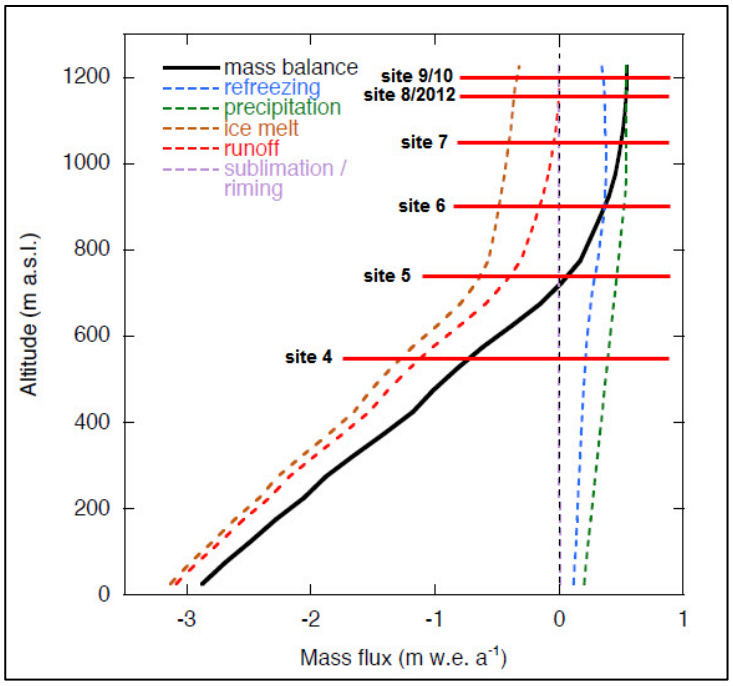

Figure 22. Components of simulated mass balance at Nordenskiöldbreen plotted against altitude. Also shown are altitudinal positions of shallow core taken in 1997 (modified from *van Pelt et al, 2012*).

The interannual variability of refreezing was found to be rather moderate (Fig. 23), which goes in line with findings of Wright (*2005*). Years with larger accumulation, lower winter temperatures and higher summer temperatures result in large volumes of internal accumulation.

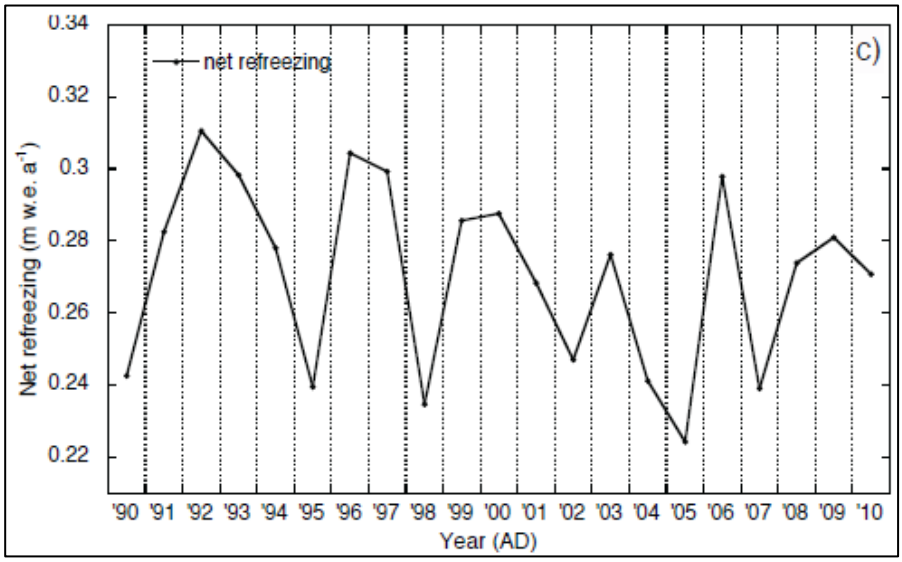

Figure 23. Simulated temporal evolution of net refreezing rates at Nordenskiöldbreen (from *van Pelt et al, 2012*).

Van Pelt et al. (*2012*) also performed experiments to identify sensitivity of mass balance components to possible change in temperature and precipitation that can be expected in a changing climate. Mass balance appears to be most sensitive to climatic variations in the lower elevation belts. The rates of refreezing appear to be not very sensitive to realistic variations of precipitation and temperature in the simulation period (2000-2100), mainly because of compensating effects.

An increase in precipitation alone will lead to increase of potential to refreezing and actual rates of refreezing provided that sufficient melt water is available, as it leads to increase of pore space and also cold content through increased mass of annual layer. Increase in temperature alone will lead firstly to higher melt rates, secondly to decrease of accumulation zone area and thirdly to decrease of the cold content of the snow/firn/ice pack. In the high accumulation zone the increased melt rates should lead to elevated rates of internal accumulation since according to the simulation the water availability is the

limitation. But the effect is likely to be temporal, since the release of latent heat effectively rises englacial temperatures. Extensive ablation zone does not favor refreezing as water runoff from the surface of bare ice goes faster and reduces the time for superimposed ice formation.

Comparing simulated refreezing rates at Nordenskiöldbreen and Lomonosovfonna with field data

Driven by apparent difficulties in direct field measurements of internal accumulation rates a number of scientists have attempted to apply the modelling approach to resolve for that component of the mass balance. Applied models vary in complexity but most credit is to be given to physically based simulations done on a detailed grid representing several internal glacier levels (*Reijmer et al., 2012*). It is of principal interest to see how well a comprehensive physically based layered englacial model is able to reproduce the englacial conditions, as this will affect the processes involved in internal accumulation and can be a measure of model accuracy in reproducing the rates of refreezing. The approach presented by Greuell and Konzelmann (*1994*) has been implemented in a number of glaciological studies, some of them aimed at estimating mass and energy balance of glaciers at Svalbard (*Wright et al., 2005; van Pelt et al., 2012; Karner et al., 2013*).

Below we present an attempt to assess the performance of SOMARS used for description of processes at Nordenskiöldbreen and Lomonosovfonna (*van Pelt et al., 2012*). Several shallow cores were drilled on Nordenskiöldbreen in 1997 by Uppsala University with collaborating institutions (*Isaksson et al, 2001*). Locations of the cores are presented in Fig. 20. The ice cores were analyzed for stratigraphy and $\delta^{18}O$. Stratigraphical logs were done following the classification of ice facies (Table 3). Each type of facies was assigned an assumed density, which was verified by empirical evidences. Density distributions derived by this type of transformation are, of course, not very precise but serve as a good first order estimate. Also a density profile measured in a core in 2012 is used for the analysis.

Table 3. Classification of snow/firn/ice strata

Description	Assumed density, kg/m^3
snow	350
coarse grained snow	410
fine grained snow	530
firn	680
coarse grained firn	745
bubbly ice	880
solid ice	910

In the figures 24-29 below we are presenting a comparison between the density profiles generated by the model for locations from where shallow cores were derived in 1997 and density distributions approximated from stratigraphical descriptions. The simulation data was kindly provided by Ward van Pelt from Utrecht University and is the same dataset as was used for drawing results and conclusions of the published study (*van Pelt et al., 2012*). The data from ice cores was kindly provided by Professor Veijo Pohjola of Uppsala University. On the left and central panels of each graph evolution of englacial temperature and density during years 1996 and 1997 is presented, in each case the red line marks the position of date 10 April 1997, which approximately corresponds to the end of accumulation season and is the time when the shallow cores were taken. The right panel shows modeled density distributions on 10 April in years 1996, 1997 and 1998 and the same data from shallow cores. For site 8 (1200 m a. s. l.) also plotted is density distribution measured in a shallow core taken in April 2012 at a nearby location 2012 (*Marchenko et al., 2012*).

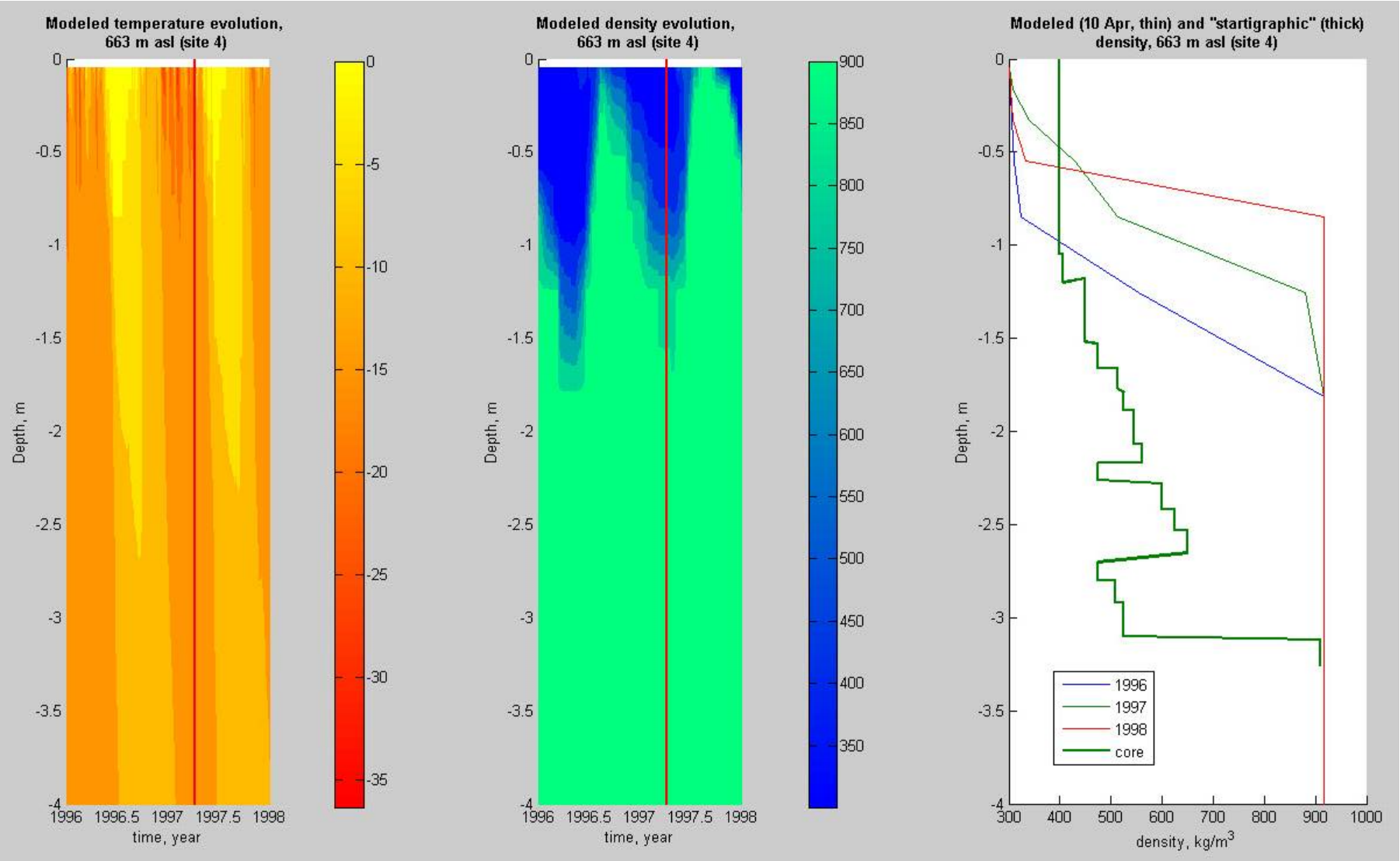

Figure 24. Shallow core location 4 – 663 m a. s. l. Evolution of englacial temperature (left panel); evolution of englacial density (central panel); comparison of modeled and observed density distributions.

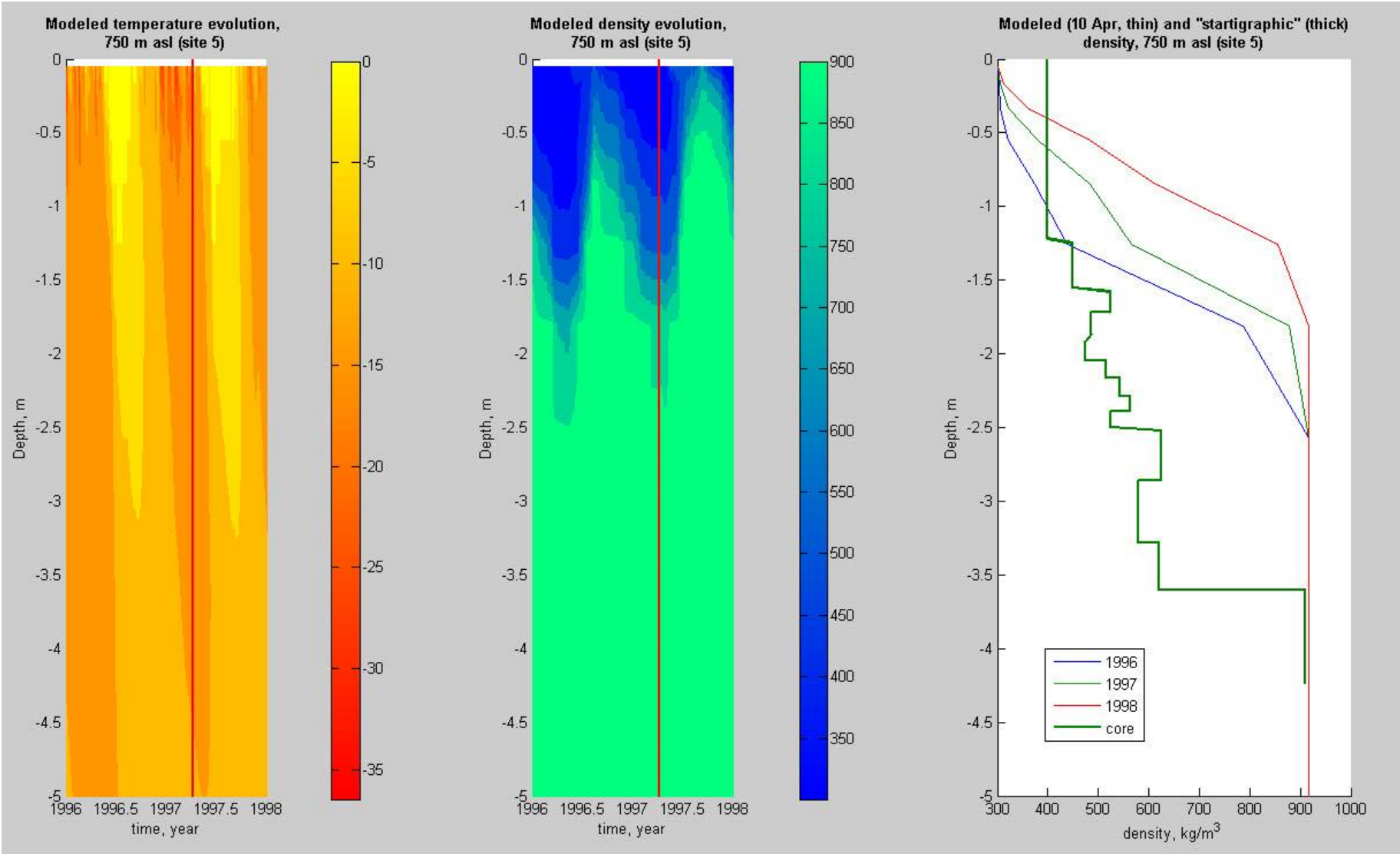

Figure 25. Shallow core location 5 – 750 m a. s. l. Evolution of englacial temperature (left panel); evolution of englacial density (central panel); comparison of modeled and observed density distributions.

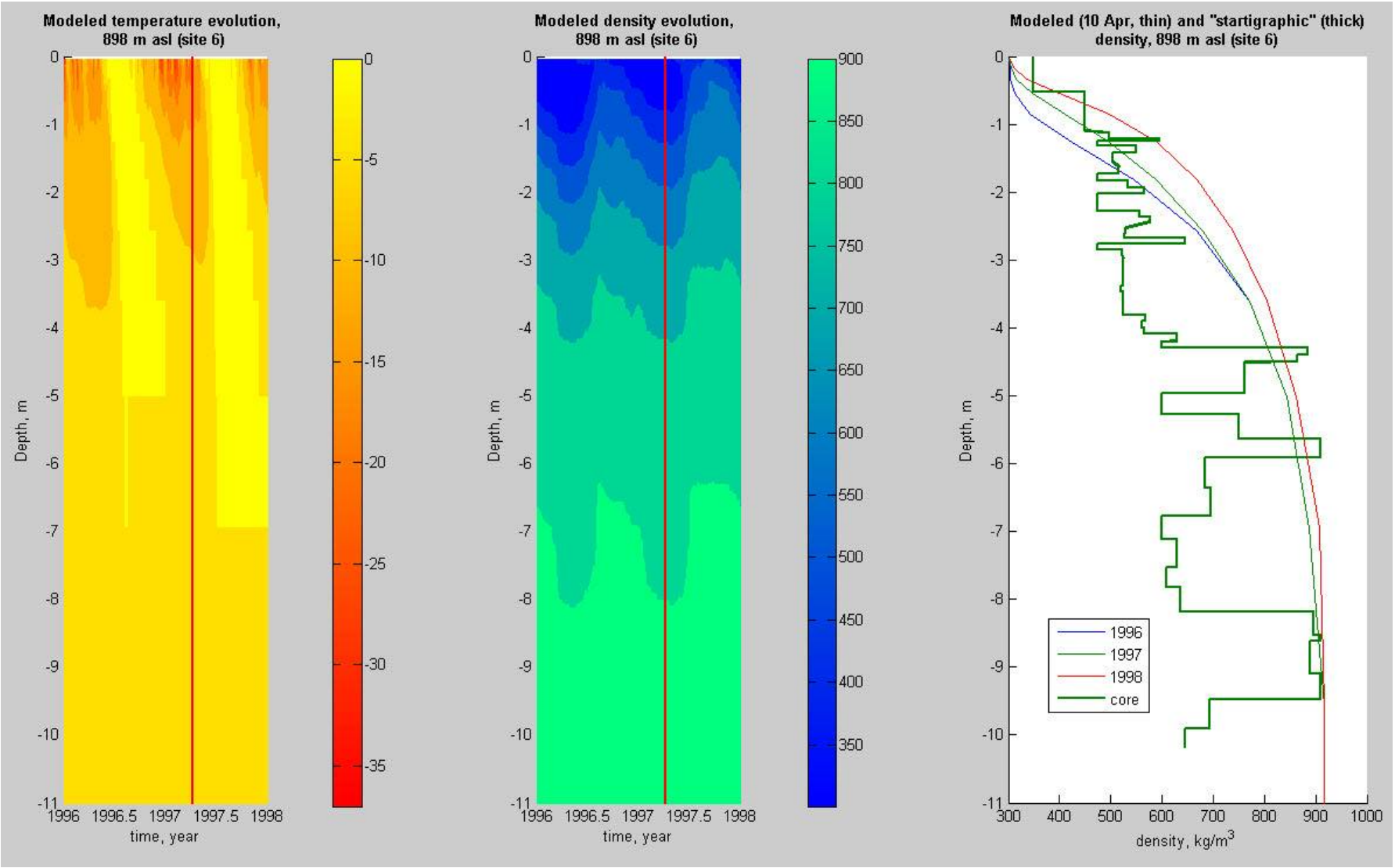

Figure 26. Shallow core location 6 – 898 m a. s. l. Evolution of englacial temperature (left panel); evolution of englacial density (central panel); comparison of modeled and observed density distributions.

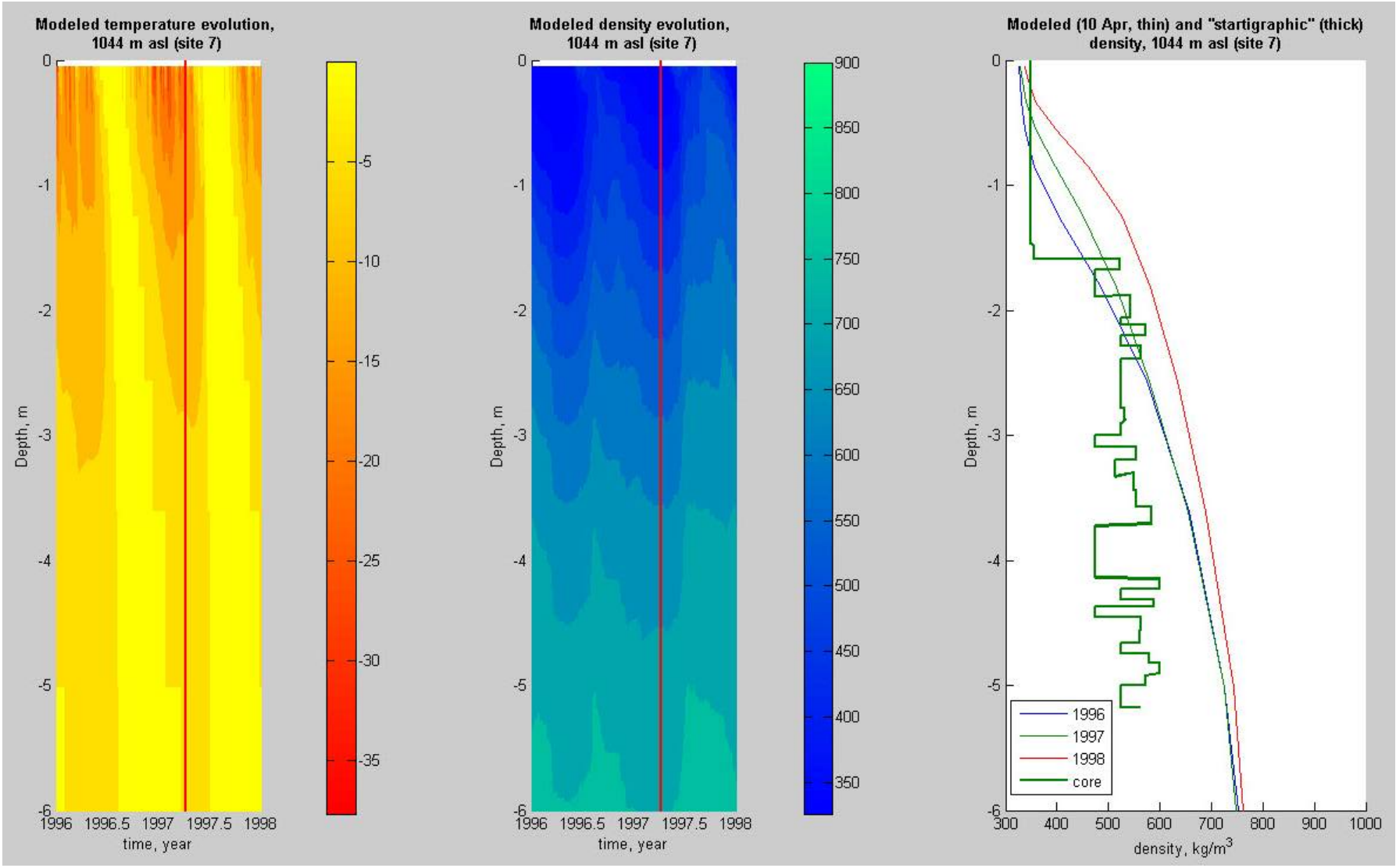

Figure 27. Shallow core location 7 – 1044 m a. s. l. Evolution of englacial temperature (left panel); evolution of englacial density (central panel); comparison of modeled and observed density distributions.

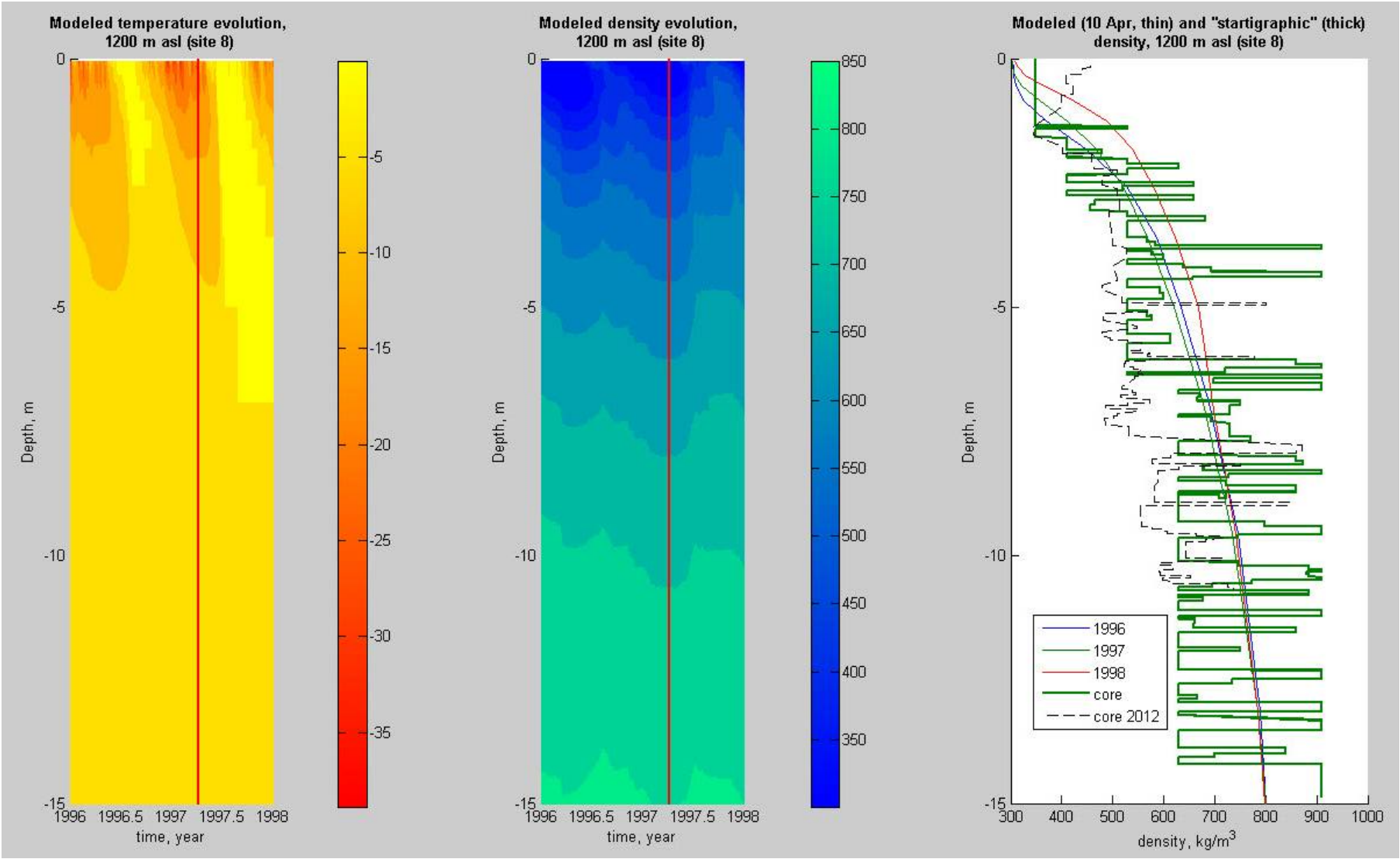

Figure 28. Shallow core location 8 – 1200 m a. s. l. Evolution of englacial temperature (left panel); evolution of englacial density (central panel); comparison of modeled and observed density distributions.

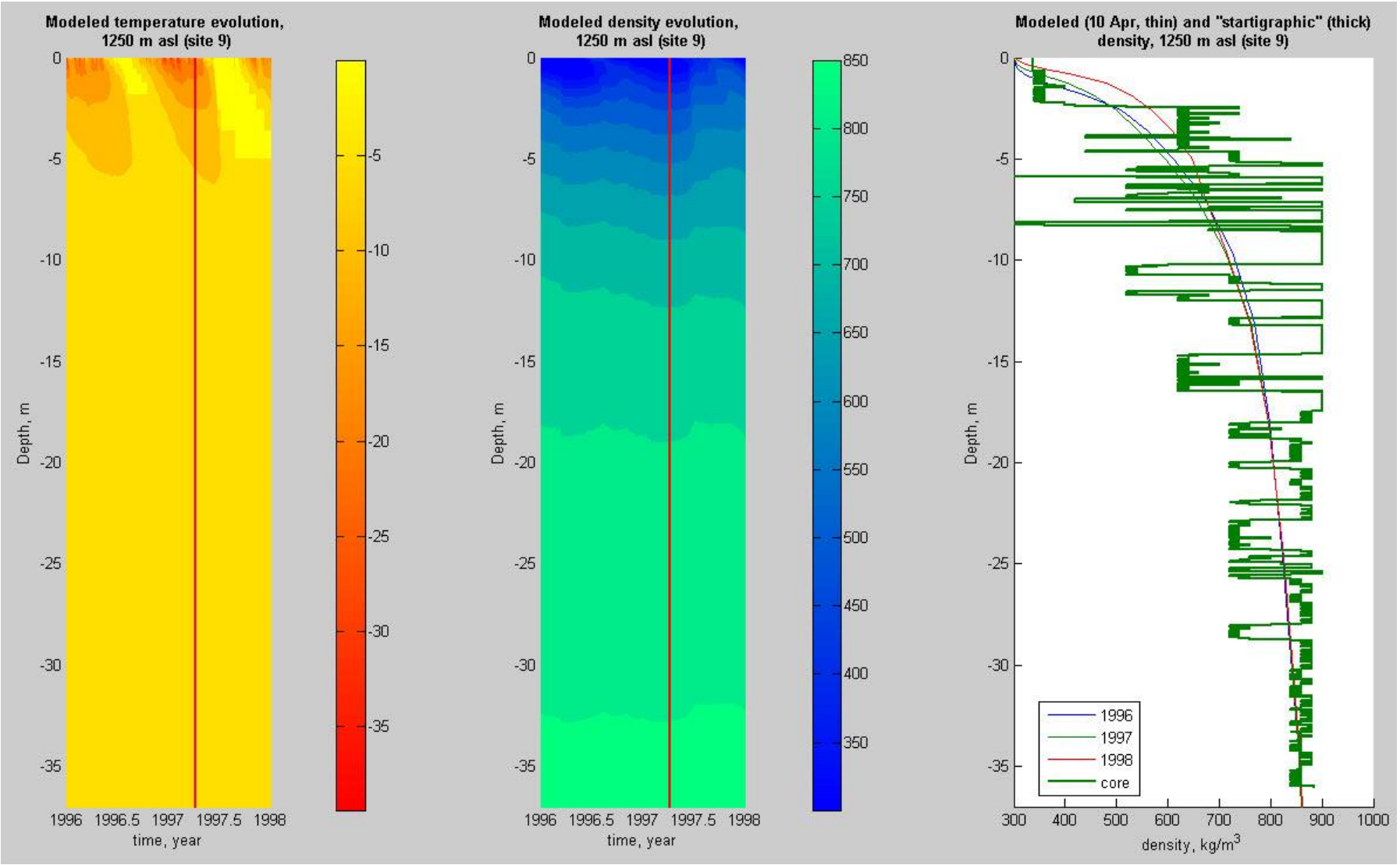

Figure 29. Shallow core location 9 – 1250 m a. s. l. Evolution of englacial temperature (left panel); evolution of englacial density (central panel); comparison of modeled and observed density distributions.

The englacial model used by van Pelt et al. (*2012*) performs well in reproducing the density profiles in the upper part of the accumulation zone, while in the lower part of the accumulation area the density is significantly overestimated. There may be several explanations for that. One possible explanation for that may be the inability of the model to generate the necessary accumulation rates which, in the long term results in refreezing of melt water in a shallower column or in other words higher concentration of refrozen mass on each meter of profile depth. Alternatively joint overestimation of cold content of snow-firn-ice pack by the end of winter and melt rates can result in higher densities. One further explanation can be in an inadequate description of melt water infiltration through the porous snow and firn pack. Indeed the model used assumes that water penetrates and refreezes down the profile as a uniform front gradually eliminating both the cold content and the pore space of snow and firn on its way. This may be not the best approximation for thick firn packs as according to observations water can create preferential flow paths and locally reach much deeper horizons than the background wetting front (*Humphrey et al, 2012*).

Results of comparison between observed and modeled density profiles are summarized in Table 4. Profile depths are limited by the lengths of the shallow cores, the other two parameters (mean density difference and mass excess) are averaged down to that depth. In the lower locations density overestimation by the model larger as there model simulates a very shallow transition from porous snow and firn to solid ice, in the cores this switch was found mush deeper. Closer to the upper reaches of the glacier the tendency is less expressed and even reversed at the top of the glacier. To get the difference between observed and modeled mass of the subsurface glacier layers we multiplied the mean difference in density by the depth of profile. The largest excess mass was produced by the model at site 6 (around 900 m a. s. l.), where the model fails to produce a thick firn pack. The apparent underestimation of mass for the highest location is perhaps misleading and is artificially exaggerated by the large profile depth, as will be shown later.

Table 4. Comparison of observed and modeled density profiles

Sites, (Elevation, m a. s. l.)	Profile depth	Mean density difference	Mass excess	
	m	kg/m^3	kg/m^2	m w. e.
4 (663)	3.26	257	838	0.84
5 (750)	4.24	186	789	0.79
6 (898)	10.18	127	1293	1.29
7 (1044)	5.18	87	451	0.45
8/2012 (1200)	14.88	8	119	0.12
9 (1250)	36.05	-23	-829	-0.83

It may also be of interest to see how the overestimation of mass in englacial profiles by the model is related to the rate of mass accumulation in each location. We first suggest expressing the thickness of profile as equivalent water layer using mean modeled density and divide it by the value of modeled mass balance in the location. This gives the approximate number of annual layers to be expected in the profile. A more reliable estimate of density overestimation by the model thus can be expressed by relation of the mass excess in the profile to the number of years therein. These results are summarized in Table 5. The maximum underestimation of internal accumulation is observed in the elevation belt of 900-1200 m a. s. l. Taking into consideration the background values of simulated refreezing rates (0.3 – 0.4 m w. e.) it can be concluded that the model reproduces the intensity of internal accumulation processes reasonably well.

The estimated error in calculated mass balance values is 20-27%. It has to be noted that uncertainties in description of processes other than refreezing also contribute to that value.

Table 5 Comparison of observed and modeled density profiles

Sites, (Elevation, m a. s. l.)	Mean modeled density along profile	Profile depth	Modeled mass balance	Number of accumulation years	Annual mass excess
	kg/m^3	m w. e.	m w. e./year		m w. e./year
5 (750)	744	3.15	0.09	35	0.02
6 (898)	764	7.78	0.37	21	0.06
7 (1044)	561	2.91	0.5	5.8	0.08
8/2012 (1200)	651	9.68	0.55	17.6	0.07
9 (1250)	745	26.85	0.54	49.7	-0.02

An estimation of internal accumulation rates can be also done on the basis of temperature measurements in glacier. A time series of englacial temperature measurements (spring 2002 to spring 2004) from Lomonosovfonna (1255 m a. s. l.) was used for estimation of refreezing rates at this glacier during one summer season. The measurements were done by Utrecht University and were kindly provided by Roderick van de Wal. The series was acquired by logging temperatures measured with 15 sensors in a thermistor string stretching down to 12.6 meters below the surface. Not all of the sensors were operational during the entire period of logging and a continuous dataset is only one year long from 11 May 2002 to 3 June 2003.

Tracking of melt water in snow and firn column by means of continuous temperature measurements proved to be powerful and efficient tool for the purpose. The mass of water refreezing in a volume of porous media is closely connected with change of the temperature of that volume because phase change results also in release of latent heat of crystallization. Another advantage of thermal tracking of melt water in snow and firn is that it does not require large scale disturbances of the snow and firn pack. The method was used in several studies for calculation of glacier mass balance (*Tsikin, 1963*), description of processes taking place in a snowpack during and after rain on snow events (*Sturm and Holmgren, 1993; Conway and Benedict, 1994*), study of water flow, stratigraphical and thermal changes in a snowpack forced by refreezing (*Pfeffer et al, 1996; Humprey et al, 2012*).

The two major processes that drive temperature evolution in the upper part of glaciers (depth 0-15 m) are conductive heat exchange and refreezing of infiltrating water. To use field measurements of englacial temperature for characterization of the latter process we need to make a correction to account for the first one. For that we run a 1-D model of temperature evolution based on Fourier's law of heat exchange:

$$\rho C \frac{\partial T}{\partial t} = k \frac{\partial^2 T}{\partial z^2},$$

where: T is temperature, t - time, z - vertical coordinate, ρ - density, C - heat capacity, k - thermal conductivity. As an initial condition temperature distribution measured before the start of melt was taken. The model is driven by boundary condition at the glacier surface being evolution of temperature measured by a thermistor at the surface. For density approximation the formulation proposed by (*Shytt, 1958*) was used: $\rho_z = \rho_i - [\rho_i - \rho_s]\exp(-C_s \cdot z)$. Tuning parameters were defined basing on data from (*Pohjola et al., 2002*): $\rho_i = 910\ kg/m^3$, $\rho_s = 500\ kg/m^3$, $C_s = 0.1$. Relative heat capacity was taken

constant at $2092.6\frac{J}{K \cdot kg}$. Heat conductivity was taken as a function of density following (*Sturm et al., 1997*): $k = 0.138 - 1.01 \cdot 10^{-3}\rho + 3.233 \cdot 10^{-6}\rho^2$.

To derive the refreezing induced temperature signal ΔT modeled temperature evolution (Fig. 30, central panel) was subtracted from the field measurements (Fig. 30, upper panel). If the observed temperature change in a grid cell is happening at a rate which is unbalanced by the modeled heat conduction at the point then the difference can be interpreted as a result of refreezing. Alternatively the reason can be in inconsistency of values for ρ, C, k_{ef}, $\partial T/\partial t$, $\partial^2 T/\partial z^2$. A comprehensive analysis of probable orders of magnitude for this type of errors is given in (*Pfeffer at al, 1996*).

To link the change of snow/firn/ice temperature with the mass of refrozen water that induced the change one can consider a snow/firn/ice volume V with mass M, having temperature T_1 and a relatively small volume of water at 0°C having mass m that is introduced inside.

The occurring process of temperature and mass change can be broken into 2 stages:

- Refreezing of water. After that M is at T_2 and m is still at 0°C. The heat energy released as a result should be equal to the change in heat content of snow/firn/ice. To describe the process we can write: $(T_2 - T_1) \cdot MC = mL$, where C - specific heat capacity of ice and L - latent heat of melt.
- Heat transfer from m to M in the result of which they both have equal temperature T_3. There should be a linear dependence between m and T_3 (if $m = 0$, then $T_3 = T_2$; if $m = M$, then $T_3 = \frac{T_2 - T_{fp}}{2}$, T_{fp} - water freezing temperature). The second process can be described by the equation: $T_3 = \frac{\left(T_{fp} - T_2\right)}{2M} m + \left(T_2 - T_{fp}\right)$

Since the relation $\frac{L}{C}$ is large we neglect the effect of the second process, this also largely facilitates calculations. Thus $\Delta T = T_2 - T_1$, where T_2 and T_1 are measured and modeled temperatures correspondingly, the mass of water that has to be refrozen to create the temperature effect ΔT can be estimated from: $m = \frac{MC \cdot \Delta T}{L}$. And the corresponding increase in density of snow and firn will be: $\Delta\rho = \rho_2 - \rho_1 = \left(M + m\right)/V - M/V = m/V$. Results are shown in Fig. 31. For estimation of internal accumulation amount we integrated m over the space domain and found the maximum value. The amount of water refreezing in the snow/firn pack that could have caused the calculated release of latent heat and increase of density is 8 cm we. This is somewhat smaller than any of the annual refreezing values produced by SOMARS model for the upper reaches of Lomonosovfonna (*van Pelt et al., 2012*). But it has to be noted that firstly the point for which we calculated the refreezing rate does not fall in the spatial domain used for simulation and is 55 m higher in altitude than the highest model node, secondly in the year 2002 one of the lowest rates of refreezing was simulated.

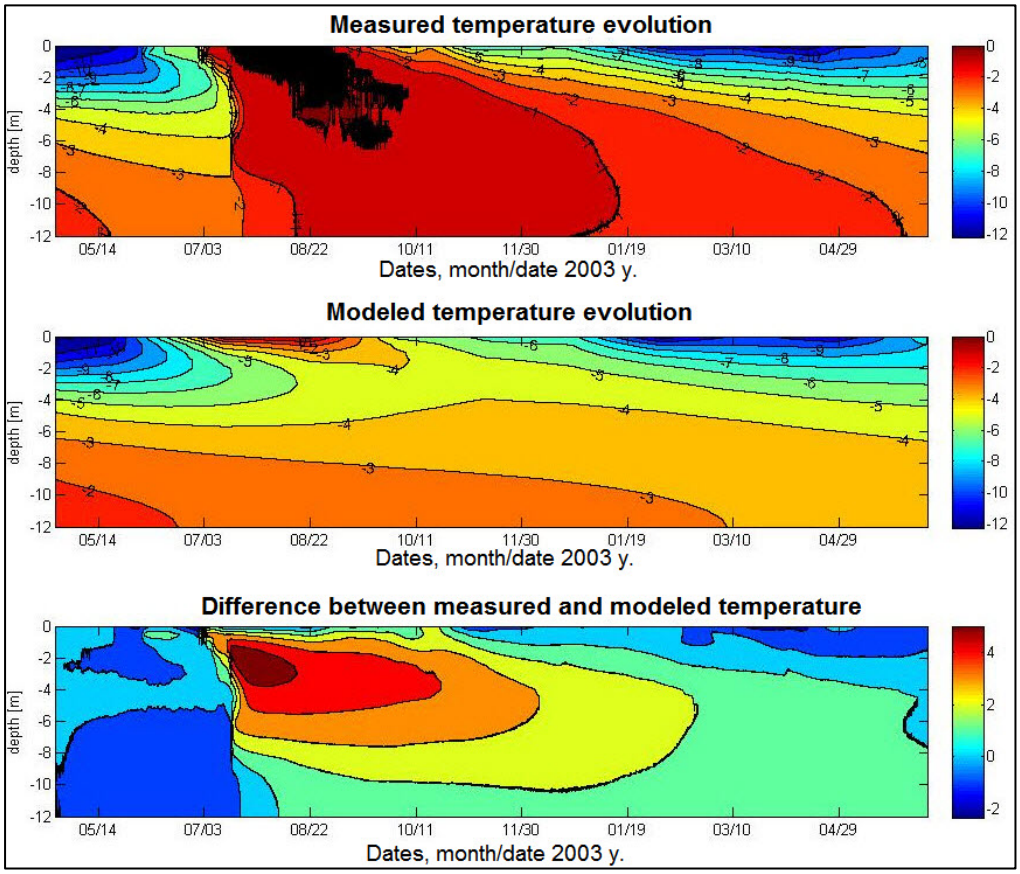

Figure 30. Upper panel: measured evolution of englacial temperature at Lomonosovfonna in 2003. Central panel: simulated temperature evolution. Lower panel: difference between measured and modeled temperatures.

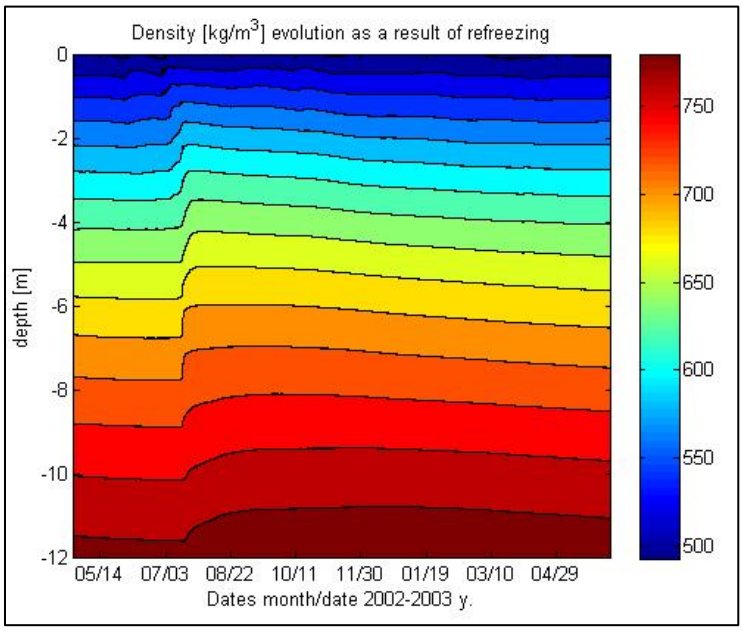

Figure 31. Calculated evolution of englacial density at Lomonosovfonna during the melt season 2003

Discussion and Conclusion

The high Arctic climate of Spitsbergen creates conditions that favor refreezing of large volumes of water inside the firn packs of glaciers and formation of superimposed ice. When combined with low rates of precipitation this source or glacier nourishment becomes very important for local and regional mass balance and can make up to 100% of its winter component. Up to date numerous efforts have been done for field measurement of internal accumulation rates. Yet, because the work demands application of non-trivial field methods and the very nature of phenomena is highly stochastic it has to be concluded that a reliable database of refreezing rates from Spitsbergen is not available up to date.

With that said it can be noted that for some glaciers, namely Midre Lovénbreen and Storöyjökulen extensive field data sets on rates of internal accumulation were presented. Both glaciers receive a substantial fraction of annual accumulation in the form of superimposed ice. Internal accumulation at Midre Lovénbreen occurs both as refreezing in pores of snow and firn and as superimposed ice and reaches 0.25 mwe/year. Storöyjökulen is almost entirely nourished by ice, that refreezes on top of the previous summer surface (*Jonsson, 1982; Jonsson and Hansson, 1990*). According to published results of stake measurements and ice core studies thickness of the layer formed on the summit of the ice cap varied from 0.2 to 0.35 m during years 1956-1980.

It can further be noted that great efforts were applied to simulate refreezing in glaciers as an integral part of their energy and mass balance. Comprehensive layered models appear to provide very good results but validation of these models is not sufficient. In the vicinity of the equilibrium line of Kongsvegen glacier up to 0.6 m of superimposed ice may be formed in one season, as it was observed in year 2000 (*Obleitner and Lehning, 2004*). But according to modelling results in most cases the ice layer (0.2-0.5 m thick) formed during spring and early summer is melted away in the course of the following ablation season. This, however, was not the case in 2001 and 2007-2010 when the study site was in the accumulation zone (*Karner et al, 2012*). Application of a simple refreezing model to simulated data on mass and energy balance of Vestfonna ice cap yielded a mean value of 0.214±0.063 m we/year for refreezing rates there (*Möller et al., 2013*).

According to modelling results (*van Pelt et al., 2012*) at Nordenskiöldbreen the rate of refreezing averaged over the spatial and temporal domains is 0.27 m w. e./year, which corresponds to 69% of accumulation rate and is just below 25% of the ablation rate. Maximum refreezing occurs at around 1000 m a. s. l., with values reaching >0.35 m w. e./year.

Parallel plots of temporal evolution of internal accumulation rates at four Svalbard glaciers for which such data is available are presented in Fig. 32. It has to be noted that for derivation of each dataset different methods were applied. Jonsson and Hansson (*1990*) reconstructed mass balance of the ice cap at Storöya island using an ice core and assumed that the glacier receives entire accumulation from superimposed ice. Other three data series are outputs of models describing refreezing that are coupled or driven by surface mass and energy balance models. While Wright (*2005*) and van Pelt et al. (*2012*) used a physically based layered model SOMARS for deriving internal accumulation rates at Midtre Lovénbreen and Nordenskiöldbreen correspondingly, Möller et al. (*2013*) applied a simplistic so called “P-max” approach for Vestfonna. The modelling results are spatially integrated over the area of the glacier in question, the data for Storöyjokulen is for the summit of the ice cap. Careful reader is referred to corresponding chapters of the current report for further details.

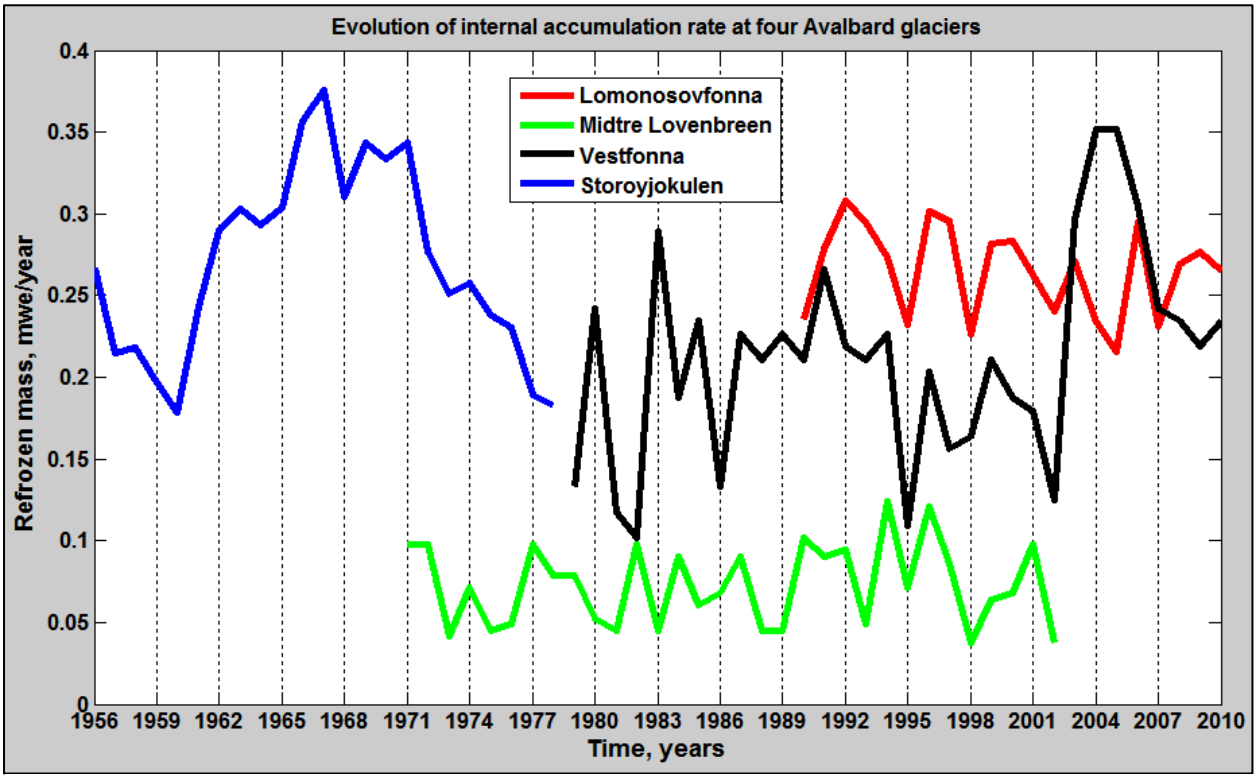

Figure 32. Evolution of rates of internal accumulation at four Svalbard glaciers.
Data: Lomonosovfonna – van Pelt et al., 2012; Midtre Lovenbreen – Wright, 2005; Vestfonna - Möller et al., 2013; Storöyjolkulen – Jonsson and Hansson, 1990.

It is apparent that the values for different glaciers lie in significantly different ranges, though large annual variability is observed. Of the four glaciers Midtre Lovenbreen exibits lowest refreezing rates, while Lomonosovfonna and summit of Storoyjökulen the highest. In certain years spikes in internal accumulation at different glaciers are matching (1995, 1996, 1998, 2002), but by no means the correlation between refreezing series can be characterized as significant.

The attempt to estimate performance of layered englacial model (SOMARS) by comparing its output for Lomonosovfonna with results from shallow ice cores allow us to conclude that it does reproduce density profile reasonably well, especially in the upper part of the accumulation zone. The model tends to overestimate densities in the lower and middle accumulation zone. The estimated error in annual mass balance can reach 20-27%.

Acknowledgments

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