Thermodynamics of the Subsurface of Glaciers with Insights from Lomonosovfonna Ice Field at Svalbard

Termodynamik under ytan hos glaciärer med inblick från isfältet Lomonosovfonna på Svalbard

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Abstract

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Glaciers are important components of the Earth's environment and are mainly found in polar and high elevation areas. They are crucial for understanding the past, ongoing and upcoming environmental changes, relevant for fresh water supply, logistical and recreational purposes. Subsurface temperature of glaciers is an important parameter heavily influencing the fluxes of mass and energy. The project focuses on how the temperature changes inside glaciers and which factors contribute to the change. Thermal conduction is one of the key processes controlling the thermodynamics of glaciers. This defines how well heat is transferred inside glaciers and how well the temperature propagates. The process of heat conduction at Lomonosovfonna ice field, Svalbard, is described using numerical simulations constrained by measured initial and boundary conditions. Simulated subsurface temperature is in line with measurements before the onset of melt in summer. After that deviations increase as the used model does not consider the process of melt water refreezing. This makes the simulation only partially successful.

Key words: temperature, glacier, refreezing, thermal conductivity, simulation

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Glaciärer är viktiga komponenter i jordens omgivning och återfinns främst i polarområden och områden på hög höjd. De är viktiga för att förstå tidigare, pågående och kommande miljöförändringar, relevanta för färskvattenförsörjning, logistiska och återskapande ändamål. Temperaturen inom glaciärer är en viktig parameter som påverkar flödena av massa och energi. Projektet fokuserar på hur temperaturen förändras inom glaciärer och vilka faktorer som bidrar till förändringen.


**Nyckelord:** temperatur, glaciär, återfrysning, värmeledningsförmåga, simulering

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1. Introduction

The aim of the project is to study the principles of glaciology with a particular focus on the processes defining the subsurface temperature in glaciers. It will result in experience of glacier thermodynamics using an existing computer code describing the process of heat conduction and temperature data from Lomonosovfonna ice field. The project will examine the feedback of the subsurface temperature to the changes of the initial and boundary conditions as well as properties of the subsurface profile. This results in the following questions of issue for the project: Is it possible to simulate the temperature of a glacier profile by using only values of initial and boundary conditions? How does the temperature change in glaciers and which processes that are involved in the heat exchange?

In cold areas at high latitudes and altitudes, glaciers are important components which are connected through various feedbacks with the hydrosphere, atmosphere and periglacial environments (Marchenko 2018). The runoff from glaciers are important for a number of different matters, being for instance agriculture, hydropower and the global sea level (Milner et al. 2017). Glaciers correspond to the prevailing climate and the atmospheric conditions at a location, hence changes of the climate and glaciers are closely linked together. They depend largely on the climate and are consequently used to reconstruct past climate changes as well as predict the corresponding between glaciers and future changes of the climate (Benn & Evans 1998). Temperature defines the dynamics of several processes and are accordingly a significant part to better understand past, present and future changes of glaciers.

The expected outcome is to obtain better understanding of which processes is involved in the heat exchange of glaciers as well as receive practical skills in describing these processes using data from measurements and computer simulations in MATLAB.

2. Background

2.1 Behaviour of Glaciers

According to Cogley et al (2011, p.45) a glacier is “A perennial mass of ice, and possibly firm and snow, originating on the land surface by the recrystallisation of snow or other forms of solid precipitation and showing evidence of past or present flow”. Glaciers are the remnants of the amount of mass of what they receive and what they lose. The extent is dependent on the accumulation of different types of frozen precipitation. Accumulation takes place when snow and ice assemble in a glacier system existing above the snow line. This is usually at the beginning of a glacier, high up in the terrain. The snow gets compacted when more snow falls,
resulting in ice at the bottom layer. Snow can accumulate in different ways, for example by precipitation of snow, by avalanches and by snow transported by winds. When snow accumulates, it gets compacted when more is added on top, leading to transformation to ice of the deeper layers and increase of volume of the glacier. This process generally occurs during winter and the opposite process, occurring in summer, is called ablation. Ablation correspond to the loss in mass of a glacier, where snow and ice disappears from the system in various ways. The loss is mainly due to melt of the glacier, sublimation and by calving of icebergs, resulting in smaller volume of the glacier in question. The sources of loss are at the surface, at the base and at the front of the glacier as well as internal loss. Because melting of the glacier leads to ablation, the process has a higher occurrence at lower elevations where it generally is warmer (Benn & Evans 1998, pp.66-69).

For a glacier to remain the same mass over a period, the mass balance of the glacier must remain the same, since it is defined as the sum of accumulation and ablation. The mass balance indicates the change of mass over a specific time interval, usually over a season or a year (Benn & Evans 1998, p.75).

Between the accumulation zone and the ablation zone a boundary can be drawn, this boundary is called the equilibrium line (Cogley et al. 2011). The equilibrium line altitude (ELA) is the altitude of which the equilibrium line is drawn (Benn & Evans 1998, pp. 82-83). One of the most important factors defining the rate of melt of a glacier and thereby also the ELA is temperature. Temperature is defined by the longwave radiation from the atmosphere being the biggest heat source of melt. The longwave radiation facilitates the transfer of air into the glacier surface, making the glacier melt and the ablation to increase, which in turn raises the ELA. After longwave radiation, sensible heat fluxes and shortwave radiation are also significant energy sources contributing to the rise of the ELA (Ohmura 2001).

Ice deforms within glaciers due to the gravity causing a constant flow movement of the ice in different directions (Souchez & Lorrain 1991). Glaciers deform under their own pressure and flow because of the pressure gradient (Benn & Evans 1998, p.242), which describes at what rate the pressure changes (Tarbuck & Lutgens 2011). All types of glaciers flow by internal deformation, where the velocity decreases with depth. If the base of the glacier is at melting point, a film of water is formed causing basal sliding. Regardless which type of movement a glacier goes through, the flow is always in the downslope-direction through sliding and deformation. This is due to that accumulation occurs in the upper part of the glacier and ablation in the lower part. If spatial imbalance between higher elevation belts occurs, it is compensated by ice flow (Benn & Evans 1998, p.242).

Depending on the rate of accumulation and ablation the glacier margin periodically advance and retreat which also is affected by the horizontal velocity component. This can be connected to the mass balance; a positive mass balance has a higher accumulation rate than ablation rate leading to advance of the glacier front. However, a negative mass balance result in more melting and a receding glacier. Even though a glacier is receding, it will still flow, although with very low velocity. A glacier will retreat if the ELA rises and the accumulation area decreases
Weather the terminus of a glacier advances or retreats can be influenced by climate changes over long periods of time (Benn & Evans 1998, p.242).

The slightest change in the climate affects a glacier’s distribution in a longer time interval. If the climate changes, glaciers change in response. Studies and monitoring of mass balance has shown that there is a certain lack of balance between the climate and glaciers. Leading to a lot of studies regarding the mass balance being done around the world (WGMS 2013).

2.2 Distribution of Glaciers on the Earth

A variety of glacier types exist on Earth, with different shapes and sizes, ranging from very small to huge ones, where the largest ones are called ice sheets and only exists on Greenland and Antarctica. The definition of an ice sheet is a large mass of ice on a continent with lateral extension due to no topographic limitation. Ice sheets cover great areas of continental size, 50 000 km² has been put as a threshold between ice sheets and smaller glaciers (Benn & Evans 1998, pp.15-16). Glaciers on the other hand are limited by the topography and exist as various types, for example valley glaciers, niche glaciers and ice fields (Benn & Evans 1998, pp.18-19). Different ways of classifying glaciers exist, based on the morphology and geophysical characteristics as well as the dynamical characteristics (Ahlmann 1948).

For glaciers to form it requires a steady supply of precipitation as well as a cold climate. The accumulation of snow and ice needs to exceed the ablation for long periods of time for glaciers to remain. This requires certain climate conditions found in mountains and polar areas. At higher altitudes, glaciers are found more frequent compared to lower land. This is partly due to that thin air cannot hold as much heat energy as denser, lower air (Benn & Evans 1998, p.39). However, the main reason is the lapse rate, meaning that the air temperature decreases when the altitude increases. This phenomenon is caused by the fact that the radiation from the sun warms up the surface of the Earth and thereby the air above. The colder air at higher altitudes makes the water vapour condense into water droplets and deposit into ice crystals, resulting in more precipitation higher up (Ahrens 2009). Except for being found frequently at high altitudes, glaciers are also more widespread closer to the poles due to the low solar angle at high latitudes, which result in less energy being available contributing to the ablation of the glaciers. The distribution is therefore found at lower altitudes closer to the poles, where their ELA is lower. Glaciers are because of this found in various parts of the world (figure 1), for example Greenland, the Rockies, the Andes, Himalaya and the polar regions (Benn & Evans 1998, p.39).

The appearance of a glacier depends on which surroundings it occurs in. The climatic conditions affect the mass balance and weather it is positive or negative the glacier have different distributions as well as different thicknesses. Climatic factors are also a big variable in the appearance. Glaciers respond to the fluctuations and changes of the climate over short as well as long time periods (Ahlmann 1953).
Figure 1. Global distribution of glaciers showed in the light blue areas. Only glaciers bigger than 5 km² are seen (Scambos et al. 2016).

2.3 Thermodynamics of Glaciers

In broad terms thermodynamics is the part of physics dealing with the relationship between temperature and heat with other types of work and energy (Buckingham 1964). There are two principle processes of heat exchange: radiation which is transport of energy by electromagnetic waves and conduction driven by temperature gradient.

In natural conditions ice temperature is much closer to the pressure-melting point (0°C to -3°C) than many other solid materials. However, the temperature of glaciers varies with time and depth within the range of 0°C to -40°C. The temperature inside a glacier is defined by the surroundings, which is how the atmospheric conditions look like at the surface and at the base. Temperature generally increases with depth due to deep processes like geothermal heat flux and basal sliding (Benn & Evans 1998, p.93). Glacier ice is not transparent to electromagnetic radiation and thus conduction becomes the dominant process of heat transfer. However, since moving glaciers are usually considered in reference to stable ground, temperature in a point is also affected by advection of colder or warmer ice downstream.
In an isolated system, energy is constant and cannot be created or destroyed. This is known as the first law of thermodynamics which states the conservation of energy, treating the parameters of change in internal energy, \( U \), amount of supplied heat, \( Q \), energy and amount of work, \( W \) (Buckingham 1964):

\[
dU = Q - W
\]  

(1)

Mass and energy flux are important parameters of thermodynamics in glaciers. Conductive heat flux, \( q \), can be described based on the parameters of thermal conductivity, \( k \), temperature, \( T \), and depth, \( z \), in the equation:

\[
q = -k \cdot \frac{dT}{dz}
\]  

(2)

The first law of thermodynamics and the definition of heat flux can be rewritten, resulting in the following equation (Cuffey & Paterson 2010, p. 401):

\[
\rho \cdot c \cdot \frac{dT}{dt} = \frac{d}{dz} \left( k \cdot \frac{dT}{dz} \right) + L + D + C
\]  

(3)

where \( T \) is temperature, \( \rho \) is density, \( z \) is depth, \( t \) is time, \( c \) is the specific heat capacity and \( k \) is the thermal conductivity. The terms \( C \), \( L \) and \( D \) describe heat flow by advection, heat production due to refreezing of water and ice flow respectively. In equation 3 the term on the left describes the change of temperature in time. The two first terms on the right describe the processes of heat flow and the two rightmost terms describe the sources of heat. Within an ice body, heat is transported in two principal ways: by thermal conduction and by advection of colder or warmer ice due to ice flow. Advection (\( C \) in equation 3) is the transport by bulk motion due to the movement of the ice (Cuffey & Paterson 2010, pp. 399).

Furthermore, there are several processes resulting in heat production. The inside of a glacier becomes warmer due to refreezing of melt and rainwater (\( L \) in equation 3) percolating to a depth where the temperature is below the freezing point. Water freezing results in internal accumulation and is accompanied by the release of the latent heat to colder surrounding ice, leading to increasing ice temperature. This refreezing of water occurs anywhere in the presence of liquid water and cold ice (Cogley et al. 2011). In ablation areas, refreezing is not effective due to runoff of the meltwater from the surface (Hagen et al. 1993).

Dynamic heating (\( D \) in equation 3) occurs due to motion of ice as the kinetic energy of ice movement is transformed to the thermal energy. Faster flowing glaciers generate more heat due to more friction. This type of energy source is only noticeable at the bed of the glacier (Benn & Evans 1998, pp. 94-95).
3. Study Site

3.1 Site Description

Svalbard holds the third most glaciers on Earth with about 6 % of the total amount excluding Antarctica and Greenland (Moholdt et al. 2010). Approximately 60 % of the surface of Svalbard is covered by glaciers of different types (Hagen et al. 1993). The project focuses on the large ice field Lomonosovfonna, located on the island of Spitsbergen in Svalbard (figure 2).

Figure 2. The location of the study site at Lomonosovfonna ice field, where “Field site” marks where the measurements are done. In the right corner there is a map of Svalbard, showing the location of Lomonosovfonna at the island of Spitsbergen (Marchenko et al. 2017).
Air temperature in Longyearbyen experiences noticeable variations in the course of a year from -15°C to 5°C (figure 3). Interannual variability in air temperature is heavily dependent on the sea ice conditions. When the sea around the island is ice-free, the atmospheric conditions become more mild and humid, whereas colder and drier winters occur when the sea ice is present around the archipelago. The latter happens because the sea ice isolate from the heat upwelling from the sea water, and the associated snow cover effectively reflects the shortwave solar radiation due to the high albedo (Øseth 2011).

**Figure 3.** The monthly average air temperature at Svalbard during the years 1961 to 1990. Data is collected from the weather station Longyearbyen. Based on air temperature values from Norwegian Meteorological Institute (2018).
As in the Arctic area the precipitation at Svalbard is usually very low compared to other locations at the latitude, with a total of about 210 mm a year (figure 4). The low precipitation rate over a year is caused by low temperature and humidity of air masses (Øseth 2011). Glaciers generally receive more precipitation than areas at lower elevations due to the orographic effect (Hagen et al. 1993).

![Figure 4](image)

**Figure 4.** The monthly average values of precipitation during the years 1961 to 1990, at Longyearbyen weather station. Based on precipitation data from Norwegian Meteorological Institute (2018).

The study site at Lomonosovfonna is found at 1200 m a.s.l. while the weather station at Longyearbyen is at sea level. Hence, altitudinal gradients in temperature and precipitation can be expected to result in generally lower temperature and higher precipitation rates at Lomonosovfonna (van Pelt et al. 2012).

### 3.2 Data Set

The data is collected by Sergey Marchenko at the Lomonosovfonna ice field accumulation zone, 1200 m a.s.l. during the period of April 2014 until April 2015. The measurements were using thermistors, which change resistivity in response to temperature (Marchenko 2018). The data set includes values of subsurface temperature and density along with corresponding times and depths. The surveyed profile in the glacier is at approximately 0.5-12.2 m below the surface in April 2014 and covers the period from April 2014 to April 2015. The relative height of the glacier surface changes however with time due to accumulation and ablation.

In figure 5 the density profile is shown along with the temperature fluctuations. The ice density varies between 900 to 910 kg/m³ while the density of snow and firn increases downwards from about 100 to 900 kg/m³ with numerous positive spikes associated with ice layers in firn. The density was measured in April 2014. The
temperature at the depth of 0.5 m in April 2014, is approximately -15°C and it gradually decreases to close to 0°C at the base of 12 m. Conductive warming occurs in May and June, while a sudden warming occurs in July because of the refreezing of surface meltwater penetrating downwards, leading to release of latent heat. After the melt season in the summer, the ice temperature gradually decreases due to the low temperature at the surface. This is illustrated in figure 5 where the temperature change in the subsurface is plotted as a function of depth and time. A warmer period of about 0°C can be seen by the yellow colour in the middle and colder periods of about -10°C at the sides.

Figure 5. The density change with depth from Lomonosovfonna is shown to the left of the figure, the depth is reference to the glacier surface in April 2014. To the right of the figure measured temperature data from April 2014 until April 2015, of the subsurface of Lomonosovfonna ice field is shown.

4. Previous Studies of the Temperature of Glaciers

4.1 Temperature of a Temperate Glacier

A report written by Harrison (1972) analyses the subsurface temperature of a temperate glacier located at Mount Olympus, Washington, U.S.A in the 1960s. The englacial temperature was measured in boreholes near the equilibrium line at the location, in an approximately 125 m deep glacier profile. Most of the boreholes were water filled at the base of the firn layer and with time they tend to close by refreezing. The temperature was estimated from the rates of downward penetration of a thermal
reamer used to maintain the holes open by melting the ice forming on the borehole walls.

The temperature at the surface was -0.03°C decreasing downwards to -0.13°C at the depth of 105 m. The results indicate that at this temperature, impurities which are water soluble are an important part of the thermal behaviour of the ice, leading to the effective heat capacity to play a greater role. It was concluded that the effective heat capacity at the glacier had a much larger value than pure ice during the same temperature.

4.2 Reconstruction of the Historical Temperature Trend from Measurements in a Medium-Length Borehole on the Lomonosovfonna Plateau, Svalbard

van de Wal et al. (2002) report high accuracy temperature measurements from Lomonosovfonna ice field done in 1997 in a 120 m deep borehole at the altitude of 1230 m a.s.l. The result indicates a close to isothermal situation at the interval of 15-120 m depth in the profile where the mean value of the temperature is -2.83°C. The glacier is cold throughout because not enough water is available in summer to warm up the part of the profile that is thermally active in the annual cycle (upper 10-12 m). In the glacier profile the general trend is increasing in temperature downwards. This increase of temperature is dependent on the geothermal heat flux warming the interior layers as well as the upward directed energy flux at the surface dragging the heat out. The maximum temperature in the profile is at 15 m depth and the minimum temperature however, excluding the glacier surface, is at a depth of a rough 70 m. This indicate a recent colder period at the glacier. When a warmer period passed, the melt increased leading to the temperature profile being moved to the right. The outside becomes warmer while the inside is still cold. However, this effect is more expressed at the surface. Similar situations have been observed at other places on Svalbard.

Based on the measured temperature profile and the modelling results authors conclude that during the last 100 years, the temperature at Lomonosovfonna has increased, which supposedly also led to an increase of the refreezing rate. The report states that refreezing was the biggest uncertainty of the temperature simulations.

4.3 Simulating Melt, Runoff and Refreezing on Nordenskiöldbreen, Svalbard, Using a Coupled Snow and Energy Balance Model

In this report, written by van Pelt et al. (2012) a study has been done regarding the evolution of mass balance and the mass budget influenced by refreezing of an outlet glacier from Lomonosovfonna, named Nordenskiöldbreen. This was done using two connected models, a distributed model of the balance of energy with a multi-layer model of snow. Meteorological data from the airport on Svalbard and data from a
regional climate model was used in the model. The model was running over a span of 21 years, from 1989 to 2010.

The results of the simulated subsurface, temperature, density and water content profiles as well as mass balance and the flux radiations are in general good correspondence with the real observations. The thermodynamic simulation is represented by two zones, an ablation zone which is cold and an accumulation zone which is temperate due to volumes of refreezing. Results indicate an overall increase in the annual mean temperature, with a few variations since 1990. The climate in the future will decrease the sensitivity of the ELA and the mass balance. It is also concluded that the change in climate do not have a big impact on the refreezing amount.

5. Method

A pre-coded function in MATLAB was used, based on a simplified equation on equation number 3:

\[
\rho \cdot c \cdot \frac{dT}{dt} = \frac{d}{dz} \left( k \cdot \frac{dT}{dz} \right) \tag{4}
\]

The equation considers the vital process of thermal conduction from a one-dimensional conductive heat flow. The processed terms in the equation is the media properties of temperature, \( T \), density, \( \rho \), specific heat capacity, \( c \), and effective thermal conductivity, \( k \), as well as the variables time, \( t \), and depth, \( z \).

Specific heat capacity is the amount of energy it takes to increase the temperature by 1°C of 1 kg of the media. It does not depend on density, but slightly on temperature. Thermal conductivity accounts for the ability of a media to transfer heat and is unlike specific heat capacity dependent on density (Cuffey & Paterson 2010, pp. 400-401). Density and effective thermal conductivity are correlated, higher density result in higher effective thermal conductivity and vice versa. The thermal conductivity data values measured in seasonal snow are scattered widely because of differences in grain sizes and grain types (Sturm et al. 1997). The wide variety is also due to the dependence on density of the snow. Hence, it is common to use a formula to calculate the right thermal conductivity during specific circumstances (Cuffey & Paterson 2010, pp. 400-401).

A research considering values of thermal conductivity was done by Sturm et al. (1997) based on 488 measurements from snow samples with the help of a needle probe. The measurements had high accuracy and the equation is used to describe how the effective thermal conductivity, \( k_{eff} \), varies with density, \( \rho \):

\[
k_{eff} = 0.138 - 1.01 \cdot \rho + 3.233 \cdot \rho^2 \tag{5}
\]

In the following text the parameterisation will be named St97.
Another examination regarding thermal conductivity has been done by Calonne et al. (2011) where the output was based on a number of three-dimensional pictures showing the microstructure of snow. The samples ranging all types of seasonal snow, however the samples were only very small in size. The only variables considered was conduction of ice as well as of the intermediate air. The thermal conductivity was discovered to increase when the density was increasing as well. It is dependent on the density of the snow and ice and results in the following equation:

\[ k_{\text{eff}} = 2.5 \cdot 10^{-6} \cdot \rho^2 - 1.23 \cdot 10^{-4} \cdot \rho + 0.024 \]  

(6)

The parameterisation will be named Ca11 in the following text.

Figure 6 illustrates the differences between the effective thermal conductivity of equation number 5 and 6. The Ca11-equation result in higher effective thermal conductivity values than the St97-equation. In this case the highest density is 900 kg/m³ and the effective thermal conductivity is approximately 1.8 W/mK.

Figure 6. A simulation of the effective thermal conductivity using equations St97 and Ca11.

A given MATLAB-code produces temperature evolution in time and depth according to the equation when given media properties, initial and boundary conditions. Forward in time and central in space finite differencing schemes where used (Recktenwald 2000). Different approximations of thermal conductivity values from equation St97 and equation Ca11 was used. To be able to receive useful density values which increases with depth a density function was used, where \( \rho \) is density, \( z \) is depth and \( \text{const} \) is a tuning parameter:

\[ \rho = 910 - (910 - 400) \cdot \exp(-z/\text{const}) \]  

(7)

The tuning parameter was set to 16 based on values from Cuffey & Paterson (2010).
To better understand the thermodynamics of subsurface layers at glaciers numerical experiments were done using both synthetic and real measured initial and boundary conditions.

6. Results

6.1 Synthetic Temperature Simulation

Figure 7 illustrates conductive heat flow forced by the temperature at the upper boundary of the domain that experiences annual cyclic change around the mean value of -30°C with the amplitude of 18°C, while the other boundary condition is stable at 0°C. In order to simulate, density equation number 7 was used with a resulting value of 909 kg/m³. The figure shows the response of the temperature during 3 years at the depth of 3, 4 and 5 m below the glacier surface.

The figure is characterised by the delay of phase, the decrease in amplitude of fluctuations and the preservation of period for greater depths. The temperature fluctuations decrease with increasing depth and the phase has been displaced with time while the temperature maximums and minimums at the different depths have the same period preserved between them.

Figure 7. A yearly cyclic change of the temperature at different depths.
6.2 Century Scale Temperature Variations in an Alpine Glacier

The illustration in figure 8 was done in order to see how the temperature changes over the depth of 100 m in an Alpine glacier. The temperature data, as well as the time and space interval, is from the Grand St. Bernard glacier in the Alps collected from a survey report done by Haeberli & Funk (1991). The surface temperature used in the simulation was scattered approximately between -13.5 to -14.6°C and plotted in a stepwise-pattern during the period of 1880 to 1980. At the base of the profile, at 100 m below the glacier surface, the temperature was as cold as about -12.9°C. Density function number 7 was used for the simulation, resulting in a value of 909 kg/m³.

As seen in figure 8, the temperature gradually increases from the surface towards the bottom, where it becomes some degrees warmer. The temperature fluctuations over the years reaches a depth of approximately 30 m, fading out with depth, while the period of the temperature fluctuations extend 20 years. Longer fluctuations in temperature reach a greater depth than shorter fluctuations. With time, the surface temperature fluctuates a few degrees, ending in a warmer period in 1980. In 1980 the warm period has the highest temperature measured at the surface during the century.

Figure 8. Temperature variations in an alpine glacier over a depth of 100 m, between the years 1880-1980.
6.3 Simulated Temperature of Lomonosovfonna Ice field

Figure 9 shows the simulated temperature of Lomonosovfonna. The simulations were based only on temperature data of the initial and boundary conditions where the temperature values in between were simulated with the MATLAB-code. The conditions were taken from the measured data from Lomonosovfonna, using equation Ca11. The depth of the profile, the period as well as the density correspond to the real measurements from Lomonosovfonna, being simulations over one year of a depth of 12 m and the density measurements from April 2014.

The temperature throughout the profile is increasing downwards to 0°C at a depth of 12 m. In the middle of figure 9 there is a warmer period with a temperature of approximately 0°C at the surface, this period is during the summer months, while the colder periods are in the winter with a temperature of about -10 to -15°C.

![Simulated temperature of Lomonosovfonna ice field.](image)
Figure 10 illustrates the difference between measured and simulated subsurface temperature at Lomonosovfonna. In order to estimate the effective thermal conductivity equation Ca11 was used.

The model fails to reproduce the temperature in the summer. In the middle of figure 10 the temperature difference is the highest, up to -5°C lower temperature in the simulated profile compared to the real measurements. However, before the onset of the melt in the summer the model reproduces the temperature rather well. The temperature difference decreases downwards in the profile as well to the sides where the difference is close to 0°C. This experiment illustrates the concept “all models are wrong, but some are useful” very well. The temperature difference decreases downwards in the profile where the difference is close to 0°C.

**Figure 10.** Temperature difference between the simulated and the measured temperature from Lomonosovfonna.
The illustration in figure 11 shows the temperature difference between the simulated temperatures at Lomonosovfonna, where equation St97 was subtracted from equation Ca11.

The highest temperature difference between the two simulations is in the summer period, where equation Ca11 results in a temperature approximately 0.8°C warmer than what equation St97 gives. In the colder period of the year the St97-equation gives a temperature up to -0.4°C below than what is given by the Ca11-equation. The resulting temperature difference from the two equations are decreasing further down in the profile to approximately 0°C at the base of 12 m.

7. Discussion

In figure 7 the temperature responses to the change in depth. Increasing depth results in decreasing of amplitude and delay of phase while the period is maintained the same. The reason for this behaviour follows from the maths in heat conduction equation number 2. Where further analyses of the equation contribute to understanding of how the temperature changes in a long-period. The reason why the phase is delayed is because the heat transfer down in the glacier profile is delayed as well, while the period remains the same because the temperature fluctuations remain one per year.

The resulted simulation in figure 8 shows that the temperature fluctuations reach a greater depth than in the illustration of the measured data in figure 5. The amplitude of temperature change at the surface is larger than at the base. However, in figure 8 the variations of the temperature reach a depth of approximately 30 m because the period of temperature fluctuations is much longer, roughly 20 years while in figure 5 the period is only one year. The amplitude is at the same time much lower in figure

![Figure 11. Difference between the simulated temperatures of Lomonosovfonna by subtracting equation St97 from equation Ca11.](image)
8, approximately 2°C compared to the annual fluctuations of tens of degrees (figure 5). In this case, the effect of longer period exceeds the effect of lower amplitudes.

A difference of the temperature can be seen when comparing the illustration of the measured data of Lomonosovfonna (figure 5) and the illustration of the simulated temperature (figure 9), as well as seen in figure 10. In figure 5 the temperature in the summer period is much higher and reaches a greater depth than in figure 9, where the temperature in the summer is not as high and the higher temperature values not as widespread. In figure 9, the simulated temperature of Lomonosovfonna shows the effect that refreezing of meltwater has on the temperature of the glacier. This is seen in the middle of figure 5, during the warmer period. The meltwater percolates from the surface down in the glacier where it refreezes and releases latent heat. This is also mentioned by Hagen et al. (1993), van Pelt et al. (2012) and Marchenko et al. (2017). However, latent heat is not considered in equation number 4, which is used in the function to simulate the temperature. By neglecting latent heat from the function, the simulated temperature does not get as high temperature values as the measured data gets in the summer period, since the supplied heat from the refreezing of the meltwater is not accounted for. However, implementation of the refreezing routine in the code would require an additional boundary condition; the rate of liquid water supply at the surface by melt and rainfall.

Using equation St97 results in a lower value of the effective thermal conductivity than using equation Ca11 (figure 6) leading to less effective conduction of the heat in the glacier. During the summer period in figure 11, the resulting difference is positive because equation St97 does not conduct heat as fast as equation Ca11. Resulting in the heat from the surface in the case of the lower $k_{\text{eff}}$-value is not able to reach the same depth at the same period as the higher $k_{\text{eff}}$-value. While in the winter period, the temperature difference is negative since the simulation with the higher $k_{\text{eff}}$-value transfer the heat away from already heated ice, while the lower $k_{\text{eff}}$-value does not transfer away the heat as fast. Because of this, equation Ca11 was considered to be better to use in the simulations of the subsurface temperature at Lomonosovfonna compared to equation St97. This was done because the equation from Sturm et al. (1997) resulted in lower values of the effective thermal conductivity than what the equation from Calonne et al. (2011) did. However, this conclusion only applies to this particular case, at Lomonosovfonna from April 2014 to April 2015. Both studies from Sturm et al. (1997) and Calonne et al. (2011) used different methods, yet both are reasonable and very thorough. They had different types of seasonal snow in account but none of them are valid since they did not consider firn. Once again, the concept “all models are wrong, but some are useful” is in the right place to say. Hence, the effective thermal conductivity estimated using equation Ca11 is probably not reliable in a thick profile of 12 m as in this case, since it was estimated only on a very small sample of snow. This can result in different values that possibly are not comparable with the reality, hence the simulated temperature differs from the true measured temperature at some parts of the profile (figure 10). Sturm et al. (1997) address that differences in grain size and grain type affect the thermal conductivity. In a grain or a
small sample of snow, the thermal properties are more or less consistent, while the properties are more likely to vary in a layer of snow or in a whole profile.

According to previous studies the subsurface temperature in glaciers has increased overall during the past century due to atmospheric warming (van de Wal et al. 2002; van Pelt et al. 2012). This depends on increase of the air temperature causing warming on the surface of glaciers. The warming produces meltwater that during refreezing, when having percolated, releases latent heat leading to warming of glaciers (van de Wal et al. 2002). The latent heat released increases the subsurface temperature leading to more melt and consequently more heating of the glacier in question. Release of latent heat had a big impact on the outcome of the figures produced in this project. By not including this parameter the resulted figures deviated from the measured data. Since the measured data from Lomonosovfonna only was measured for one year, an increase of the subsurface temperature with time could not be observed.

According to the results compiled by van de Wal et al. (2002) the surveyed glacier was cold throughout compared to the provided data set from Lomonosovfonna ice field where the temperature reached 0°C after 10 m. The cold glacier did not have enough water available in summer to warm the upper 10-12 m, which indicate that the study site of Lomonosovfonna in this report received enough water in summer to warm the inside. The study site is at a different location at Lomonosovfonna than the one in the report by van de Wal et al. (2002).Nevertheless, the altitude of the study site of this project is 30 m lower in altitude, resulting in warmer surroundings and more meltwater which in turn result in higher temperature close to the surface. The results from the simulations support the assumption of van de Wal et al. (2002), who missed the refreezing routine in the reconstructions of the conditions at Lomonosovfonna due to a hard time assessing it. The simulations in this project show direct measure of by how much a pure conduction model can fail.

The temperature at Lomonosovfonna was discovered to increase with depth while the measurements from the temperate glacier indicate a decrease in temperature further away from the surface (Harrison 1972). Decreasing of temperature is an odd behaviour of the temperature profile of a glacier. The range of the temperature values of the temperate glacier, 0.03°C to -0.13°C, indicate a different situation compared to Lomonosovfonna where the range is 0°C to -15°C. The fact that the temperature is decreasing with depth is due to the surrounding conditions at the surface and at the base. The temperature inside glaciers are dependent on many factors, for instance geothermal heat flux and basal sliding, which is addressed by Benn & Evans (1993). These factors could not have been significant in the surroundings of the temperate glacier because of the decreasing temperature.

Climate changes and the impact on glaciers can be observed in these types of studies and thus be used to reconstruct past climate changes. A variation of the temperature in the surroundings can be observed in the glacier studied by van de Wal et al. (2002), hence it is safe to say that the glacier was exposed to a colder period. This way it is also possible to predict future climate changes and how the climate corresponds to glaciers.
In this project the sources of errors can be divided into measurement errors and simulation errors. Any measurements are uncertain and there are always errors in some extent. The temperature data contains several kinds of errors, errors in the measurement of temperature as well as the depth measurements. There can be errors resulting from postprocessing of the measurements. Errors can result from measurements or approximations of the density as well as in the estimates of other parameters as specific heat capacity and effective thermal conductivity. The studies regarding thermal conductivity by Sturm et al. (1997) and Calonne et al. (2011) did not consider firn, meaning that resulting values of thermal conductivity are not reliable in this simulation.

In the simulations there are errors resulting from the imperfection of the conceptual model. The temperature results relay on a model describing the thermal conduction in one vertical dimension. However, heat fluxes can be horizontal as well and if this is the case at Lomonosovfonna they will not be captured in the model. The horizontal heat flow becomes relevant in autumn when water suspended in pores after the melt season freezes. The model that was used does not describe any processes related to liquid water resulting in a significant error in the simulations. In this case the other neglected parameters in equation 3, dynamic heat and advection, would be negligibly small or not valid and thus would not affect the resulting temperature simulations. There are always numerical errors in simulations since continuous field of temperature changing in time and space is approximated by finite differences.

8. Conclusions

The temperature inside glaciers is dependent on the surroundings of the surface and the base. The atmospheric conditions, the glacier flow as well as the geothermal heat flux have huge impact on the subsurface temperature. The balance between all heat sources determines the actual temperature inside glaciers and whether it increases or decreases with depth. The temperature usually increases with depth at different rates resulting in different basal temperatures. Results show that refreezing of meltwater which releases latent heat to a large extent define the subsurface temperature. Simulating the subsurface temperature of a glacier profile with values only from the initial and boundary conditions was only partly succeeded. If latent heat would have been considered in the function used for the simulations, the result may have showed more consistent results with the measured data from Lomonosovfonna in the summer period. Using this model to reproduce temperature is very useful during winter and spring season while it is inadequate during summer in these types of environments. Hence, the method would be more useful in cold environments where no melt occurs during summer.
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References


Internet Resources

