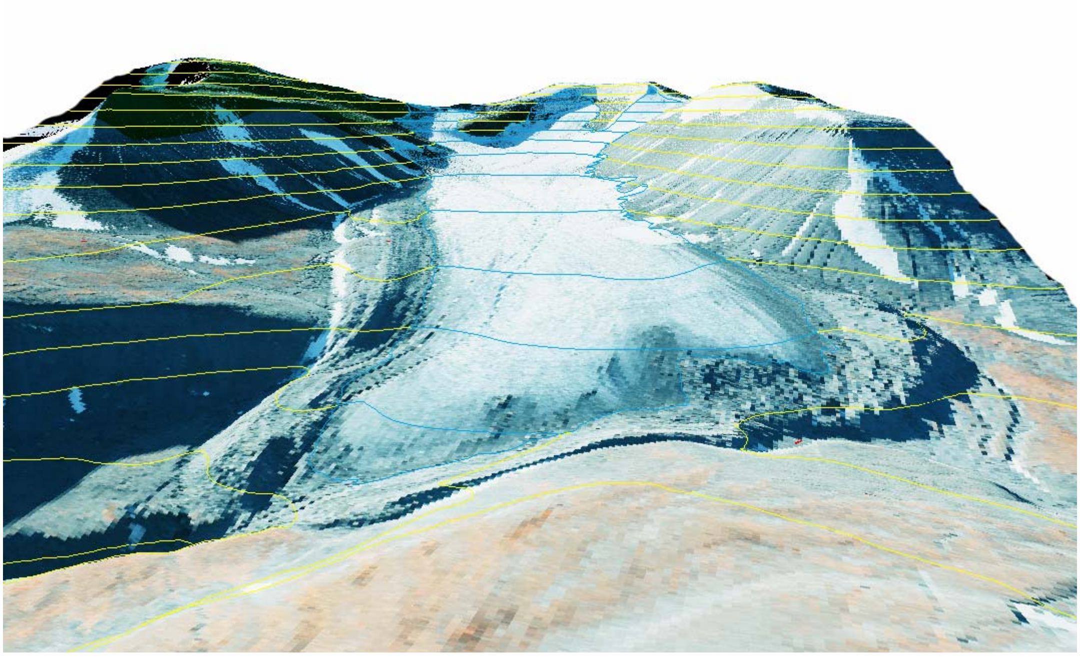
Climate – glacier links on Bogerbreen, Svalbard

Glacier mass balance investigations in central Spitsbergen 2004 / 2005

Ullrich Neumann



UNIVERSITY OF OSLO
FACULTY OF MATHEMATICS AND NATURAL SCIENCES

UNIS

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Department of Geosciences
Faculty of Mathematics and Natural Sciences

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Frontpage photo: A mosaic of 6 orthophotographs, generated from the 1990 aerial survey of Bogerbreen, draped over a digital elevation model. Heights information obtained from Norwegian Polar Institute.

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Abstract

Glaciers are key indicators for climate change. Mass balance studies form the important link between advances and retreats of glaciers to changes in climate. Mass balance studies were performed in the balance year 2004/05 on Bogerbreen as part of my Master thesis. Bogerbreen is a valley glacier with a size of 3.3 km² located in central Spitsbergen, Svalbard at 78 degrees north and 15 degrees east. The direct glaciological method was applied to measure winter mass balance using snow soundings and snow pit studies. Summer balance was obtained from stake readings. The use of artificial reference horizons and shallow ice cores helped to estimate superimposed ice formation. In addition to the primary mass balance components, air temperatures were recorded on site as well as daily glacier photographs were collected, stake and glacier surface surveys performed. Ice surface velocities were measured and I generated a high resolution digital terrain model and an orthophoto of the study site. Winter balance 2004/05 was calculated to be 0.55 ± 0.08 m water equivalent. Redistribution of snow is a major control on the spatial distribution of the winter mass balance. In a more regional scale it is likely that wind drifted snow is deposited on Bogerbreen and adds significantly to the winter mass balance beside winter precipitation. Summer mass balance were calculated to - 1.35 ± 0.24 m water equivalent. Ablation decreases with increase in elevation most likely due to the influence of turbulent heat fluxes. At the middle part of the glacier, albedo effects and shading lead to a distinctive lateral summer balance asymmetry. It is likely that the spatial distribution of winter snow plays a major role causing this lateral asymmetry. The resulting net mass balance was negative in the balance year 2004/05, with - 0.80 ± 0.20 m water equivalent. Mean geodetic net balances for the period 1990 to 2003 and previous direct measured balances indicate that Bogerbreen has not been in balance with the climate for the last decades. This is in accordance with findings elsewhere on Spitsbergen.

Content

1. Introduction

1.1 Background

About 10 % of the earth land surface area covered with glacier ice at the present (Patterson, 1994). Beside, glaciers are astonishing and breathtaking features of our nature. Interaction between glaciers and human activity exists in many ways. Studying glaciers serves foremost the purpose to answer questions concerning: management of regional water supplies in glaciated catchments, glacier related hazards, the glaciers contribution to sea level changes and past to future climate scenarios and last but not least to get a better understanding of glaciers itself (Kaser *et al.*, 2003).

Studies concerning changes in the mass of a glacier and their distribution of over space and time are termed glacier mass balance studies. The change of glacier mass over a year in particular, forms the important link between events connecting advances and retreats of glaciers with changes in climate (Patterson, 1994). Even by knowing that glaciers react in a complex manner to climatic variations (Kaser *et al.*, 2003), glaciers are key indicators for climate change (Østrem and Brugman, 1991).

Compared to other large ice masses such as Antarctica and Greenland, the Arctic hosts about two thirds of the Earth’s small glaciers. The volume of water stored equals a rise in global sea level of 0,5 m (Meier, 1984). Despite the fact that the arctic inherits only a small amount of the overall ice stored on land, it plays a major roll within the overall balance of ice masses. Due to the fact that many ice masses of the arctic show temperate or polythermal temperature regimes, they are expected to respond more rapid to climate warming compared to ice sheets of Antarctica and Greenland. A better knowledge of ice masses in the Arctic helps to delimitate uncertainties for identifying causes for past or future sea level changes (Krimmel, 1999; Dowdeswell and Dowdeswell, 1997; Jania and Hagen, 1996; Meier, 1969).

The archipelago of Svalbard is located within 76° and 80° N latitude and 10° and 33° E longitude at the north-western part of the Barents Sea shelf. The biggest of the islands is Spitsbergen (38.000 km²). The total land surface extends to 63.000 km² while at present 60

% of the land area is covered by glaciers (Hagen, 1993; Hjelle, 1993). Svalbard has a magnificent landscape. Mountains with elevations up to 1700 m show partly alpine, partly plateau character, and are often cut by wide valleys which extends into the sea. Strandflats as well as fjord landscapes are most common along the coastline of Svalbard (Hjelle, 1993). The archipelago inhabits many common arctic species of fauna and Flora. Together with the beauty of the landscapes this makes it to a very fascinating place on earth (fig. 1.1).

Figure 1.1 The location of the archipelago of Svalbard in the northern hemisphere. From (Geo-NP_NETT, 2006).

1.2 Objective and scope

The thesis is placed within the framework of a newly set up mass balance programme which started in 2004 as a joint effort of University of Oslo, University Centre in Svalbard (UNIS) and the Norwegian Polar Institute. A main objective of the study is to investigate important key parameters concerning glacier mass balance. In this context it foremost serves the purpose of, improving our fragmentary knowledge about the links between glacier responses to climate change in central Spitsbergen.

1. Introduction

In a broad sense the aim of the study is to determine the glacier – climate relationship on Bogerbreen by using mass balance investigations. Seen in a somewhat closer context my goal is to closer investigate the related variables which control mass balance on Bogerbreen and how they are linked to climate parameters by applying an appropriate method. The work presented here will focus on the following questions:

- What is the mass balance of Bogerbreen in the balance year 2004-05?
- What are the controls of the mass balance of Bogerbreen?
- What is the spatial distribution of the mass balance parameters on Bogerbreen?
- Is the glacier in balance with today’s climate?
- Does Bogerbreens mass balance follow the trend observed in other parts of Svalbard?

The following factors where decisive for choosing Bogerbreen as the study site:

- The importance of Bogerbreen’s melt water production as a contributor to the nearby settlements water supply.
- Previous published mass balance data in the period 1974-1986 is available for analysis.
- The glacier is representative in its size for the glaciers apparent in the central region of Nordenskiöld Land despite the fact that is only little representative to the whole of Svalbard due to its comparable small size of approximately 5 km².
- The close neighborhood of the field site to the settlement of Longyearbyen and UNIS ensures comparable simple logistic and easy access.

1.3 General definitions

1.3.1 Mass balance definitions

The below given definitions are in general use (Benn and Evans, 1998; Patterson, 1994; Østrem and Brugman, 1991). As early scientific research started in the late 19th century many definitions has been adapted to glacier systems in the middle latitudes and problems might arise by applying the definitions to glaciers in higher latitudes (Patterson, 1994).

Accumulation includes all processes that increases the glaciers mass and include material added as snow, ice or rain or due to avalanching, rime formation, and snow redistribution whereas *ablation* includes all those processes that remove mass. Examples are run-off, evaporation, snow redistribution and calving. The *glacier mass balance* at each point of the glacier is measured relative to the last year's *summer surface*. In order to retrieve the mass balance of the entire glacier total values are integrated by using a sufficient number of point measurements. The *balance* (b) is the change in mass measured at a point at any time. The result can be either negative or positive. One *balance year* can be defined by the time span between two successive minima (summer surfaces) of the glacier thickness at a given point. The so defined balance year can also be termed as the *stratigraphic system* and will seldom equal a calendar year. In another case the balance year might be defined in a "*fixed-date system*". Thereby the dates of the successive balance measurements must be stated. The maximum balance value during a balance year is called the *winter balance* b_w. The first part of the balance year generally shows an increasing trend for glaciers at higher latitudes. The time until the maximum is reached (i.e., the end of the accumulation period) is termed *winter season* and will be followed up by the *summer season*. The *summer balance* b_s represents changes of mass during summer season. The *net balance* b_n at a given point on the glacier is the change in balance during one balance year. It can be expressed in equation (1.1) as the algebraic sum of the winter balance and the summer balance. The values for the balances are given in meters of water equivalent.

$$b_n = b_w + b_s \qquad \textbf{(1.1)}$$

The glacier can be divided into an *accumulation area* where $b_n > 0$ and an *ablation area* where $b_n < 0$. The *equilibrium line* marks the boundary between the two. In most cases, processes changing the mass balance of glaciers occur *supraglacially* within a thin surface layer. Nevertheless also *englacial* processes within the glacier and *subglacial* beneath the glacier can play a significant role. If snow survives at least one year on the glacier but not metamorphosed to polycrystalline glacier ice, it is called *firn*. Depending upon the temperature conditions within the firn, *internal accumulation* can occur if water percolates through the firn and refreeze below the summer surface. Another process of accumulation is the formation of *superimposed ice*. It forms by refreezing of water on top of cold impermeable ice below (Hagen *et al.*, 2003a). Other then glacier balances concerning points, balances referring to an area of a glacier use the same terms but they are symbolized by using capital letters instead (B_n, B_w, B_s). The most useful parameter for summarizing the change in a glacier in a given year is the *mean specific mass balance* $\overline{b_n} = B_n / S$ where S is the area of the glacier. If $\overline{b_n}$ are calculated for individual altitude intervals and plotted against altitude, the then established plot shows the *vertical mass balance profile* (VBP) of a glacier. Even mass balance will be not the same from year to year, the shape of the curve will typically remain unchanged (Meier and Tangborn, 1965). The VBP characterizes the climate-glacier regime.

1.3.2 Mass balance methods

It is possible to determine the mass balance by using a variety of methods. Namely direct- and indirect glaciological method, geodetic method, hydrological method, flux method and modeling method. Throughout this study only the direct glaciological method and the geodetic method will be applied and therefore it is given more emphasize here. A short and comprehensive summary of the two methods is presented here from (Kaser *et al.*, 2003; Patterson, 1994).

Also called the direct method it is the only one based on *in situ* measurements. At a representative number of individual points, the change in surface elevation is measured between two dates. By multiplying the near surface density with the change in surface, we achieve an estimate of the mass balance at that point. The change in surface elevation can be measured using stakes drilled into the glacier, and snow depth investigations relative to a known stratigraphic surface such as the previous summer surface. While density values for ice are assumed to be constant, the density of snow is measured in a representative number of snow pits dug down to a reference surface or by coring. The most accurate method to date is using a contour map of net mass balance and the total mass balance B_n is calculated from equation (1.2) in its discrete form,

$$B_n = \sum_{1-n} b_n \Delta S_n \qquad \textbf{(1.2)}$$

where ΔS_n is the area of the glacier over which the net balance b_n applies.

The geodetic method establishes the values for a volume change of the glacier by subtracting the surface elevation over the entire glacier extent at two different times. The mass balance can be converted from the volume change by estimating the surface densities of the glacier. The geodetic method utilizes topographic maps or digital elevation models (DEM) derived from ground based or remotely sensed survey techniques.

1.4 Previous work

1.4.1 Mass balance studies in Svalbard

The first systematic mass balance study started as early as 1950 on Finsterwalderbreen in south west Spitsbergen. This was carried out by the Norwegian Polar Institute. Field visits were undertaken every other year until 1966. Therefore mass balance data is only given as mean values for every other year. Again in 1966 mass balance measurement started on Brøggerbreen and in 1967 on Lovénbreen. Both glaciers are located on Brøggerhaløya in the north western part of Spitsbergen. Mass balance measurements has been carried out since then on an annual basis using the direct glaciological method for accumulation and ablation measurements (Hagen, 1993). Brøggerbreen has therefore the longest continuous mass balance record on Svalbard.

In the same period, starting in 1966 Russian scientists initiated systematic annual mass balance measurements on Vøringbreen in the Grønfjord area, in the central western part of Spitsbergen. This lasted until 1991. Measurements on Bogerbreen lasted from 1974 until 1986 and 1976 until 1982 on Longyearbreen under Russian leadership. Both glaciers are located in central Spitsbergen. Mass balance measurements on Bogerbreen were reinitiated by Norwegian scientists in 2004 and continue on an annual basis. Bertilbreen in central Spitsbergen was firstly measured in 1975 and mass balance investigation was carried out until 1985 as a programme under Russian leadership. The scientists chose also a glacier situated in central eastern Spitsbergen to perform mass balance investigations on Daudbreen during the period from 1978 – 1983 (Jania and Hagen, 1996; Kotlyakov, 1985). Mass balance studies performed on Hansbreen in Hornsund, southern Spitsbergen by Polish researchers started in 1989 and is ongoing. Locations of mass balance measurements are given in figure 1.2.

Figure 1.2 Archipelago of Svalbard excluding Bjørnøya to the south. Black squares indicate the location of glaciers with direct mass balance measurements while circles displays glaciers with ice core measurements. Land area in white indicates glacier cover whereas gray shaded areas correspond to land surface non glaciericed at present day. From (Hagen *et al.*, 2003a).

Apart from the Polish mass balance program on Hansbreen, the glaciers on which mass balance investigations are performed, are relatively small (2-6 km²) and mostly below 500 m a.s.l.. Those, mostly isolated cirque or valley glaciers were only little representative for the whole of Svalbard. Until 1987, when the mass balance program on Kongsvegen in the north western part of Spitsbergen were started, only sporadic measurements were performed on larger ice masses. Especially in central and eastern Spitsbergen almost no mass balance data is available. There, investigations are hampered by high the need of high logistic efforts. In wider terms mass balance measurements have been performed on several other glaciers over the whole archipelago mostly to determine mean net balance by using a variety of different methods. An overview of mass balance investigations done on an annual basis for a period of at least three years is given in table 1.1.

Table 1.1 Containing glacier name, surface area and the periods where mass balance measurements were undertaken. (Table modified from (Jania and Hagen, 1996; Hagen, 1993).

Glacier name	Area (km²)	Annual mass balance investigations (period)
Finsterwalderbreen	44.50	1950 – 1966 (every 2nd year)
Austre Brøggerbreen	11.80	1967 - present
Midre Lovénbreen	5.95	1978 – present
Vøringbreen	2.10	1974 - 1991
Bogerbreen	5.20	1975 – 1986 and 2004 - present
Bertilbreen	5.40	1975 – 1985
Longyearbreen	4.00	1977 – 1982
Daudbreen	6.00	1978 - 1983
Kongsvegen	102.00	1987 – present
Grønfjordbreen	38.30	1988 – 1991
Fridtjovbreen	48.70	1986 - 1991
Hansbreen	57.00	1989 – present
Linnèbreen	3.85	2004 - present

1.4.2 Mass balance studies on Bogerbreen

One reason for starting mass balance investigations on Bogerbreen was to collect data from more centrally located glaciers in Spitsbergen. On the other hand, the “western-minded” settlement of Longyearbyen in close range and the then present political conditions, made Bogerbreen also to an attractive study site (Dr. Serguei M. Arkhippov, pers. comm.). Net mass balances are available for the entire period from balance year 1974/75 until 1985/86. Winter and summer balances have been measured except for the balance years 1975/76 and 1985/86 where there are only net balances available from (Troitskiy, 1989; Gokhmann *et al.*, 1988; Guskov and Troitskiy, 1987; Kotlyakov, 1985; Guskov and Troitskiy, 1985) (table 1.2). For three selected balance years 1982/83, 1983/84 and 1984/85 contour maps were published for specific winter balance and specific net balance from (Gokhmann *et al.*, 1988; Guskov and Troitskiy, 1987; Guskov and Troitskiy, 1985). Difficulties in interpreting this data arise through the fact that papers where mass balance data is presented are mainly published in Russian language.

Table 1.2 Bogerbreen mass balance series obtained between 1974 and 1986. Modified from (Troitskiy, 1989; Gokhmann *et al.*, 1988; Guskov and Troitskiy, 1987; Kotlyakov, 1985; Guskov and Troitskiy, 1985).

Balance year	Winter balance [g cm^{-2}]	Summer balance [g cm^{-2}]	Net balance [g cm^{-2}]
1974/75	57	57	0
1975/76	-	-	-20
1976/77	62	88	-26
1977/78	34	115	-81
1978/79	61	168	-107
1979/80	48	113	-65
1980/81	56	92	-36
1981/82	38	13	+25
1982/83	48	78	-30
1983/84	62	123	-61
1984/85	54	111	-57
1985/86	-	-	-60

Ice penetrating radar studies have been undertaken in the period 1975 and 1977-1980 (Macheret *et al.*, 1985). Maps displaying the subglacial relief, ice thicknesses and location of the radar profiles on Bogerbreen were obtained by air based ice penetrating radar (fig. 1.3). The literature does not clarify in which year the radar investigations are performed on Bogerbreen.

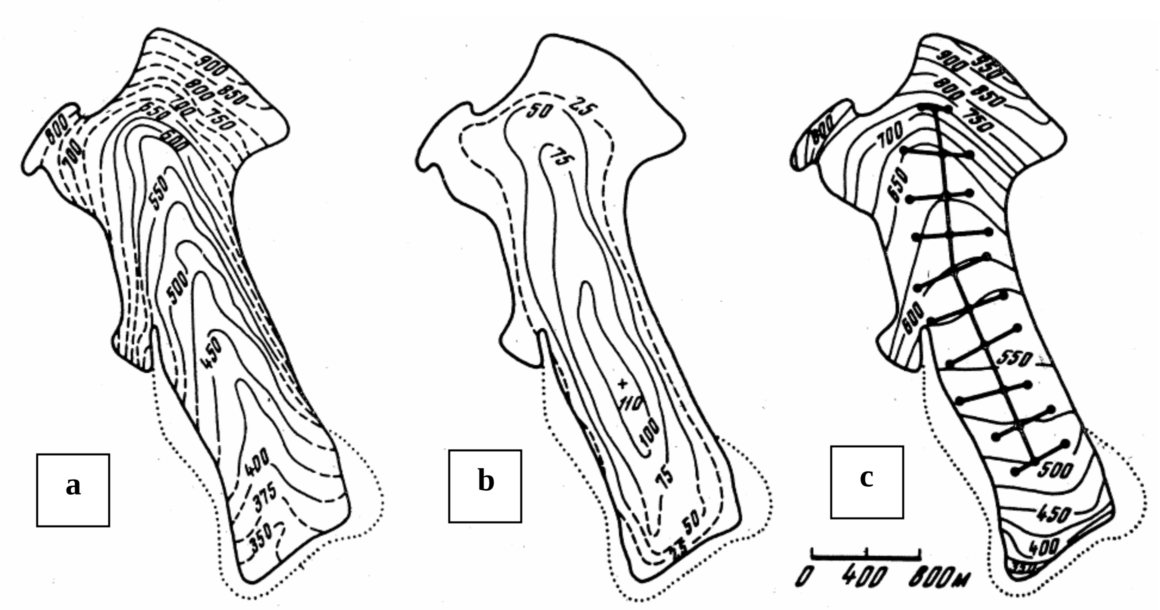

Figure 1.3 Subglacial relief (**a**), ice thicknesses (**b**) and location of the radar profile (**c**) on Bogerbreen obtained from air based ice penetrating radar between 1975 and 1980. From (Macheret *et al.*, 1985).

2. Study area

2.1 Geographical setting

The Bogerbreen field site and its surroundings is located at latitude of 78° 080 and 15° 380 longitude. The glacier is situated in a central part of Spitsbergen in the basin of Nordenskiöld Land (fig. 2.1). The basin is characteristic for its little glacier cover of below 30 % (Hagen, 1993). There, due to low precipitation, glaciers tend to be smaller compared with coastal areas in the west and east (Humlum, 2002). The Bogerbreen catchment is embedded in Karl Bayfjellet (1015 m.a.s.l.), Westbytoppane (980 m.a.s.l.) and Bingtoppen (910 m.a.s.l.) in the east, south and west direction respectively. Bogerbreen’s terminus is situated right at the drainage divide between Endalen catchment and Fardalen catchment. Surface drainage from Bogerbreen is directed into both catchments. The geology of the area is characterized by Tertiary, flat-lying sedimentary bedrock, mainly shale, silt- and sandstone dominates the lithology of the study area. Mechanically, the bedrock is soft and fine-grained (Major and Nagy, 1972). In most areas on Svalbard, ground temperatures are below 0 °C and permafrost depths vary from 200 m to 450 m in the mountains, known from mining activity and borehole measurements (Isaksen *et al.*, 2000; Liestol, 1977). Closest distance to the sea is about 12 kilometers in north-west direction towards Isfjorden. The distance to the open sea is of about 50 kilometers towards the west into the Fram Street and 70 kilometers towards the east into the Bering Sea. Svalbards biggest settlement Longyearbyen is located only 10 km from Bogerbreen.

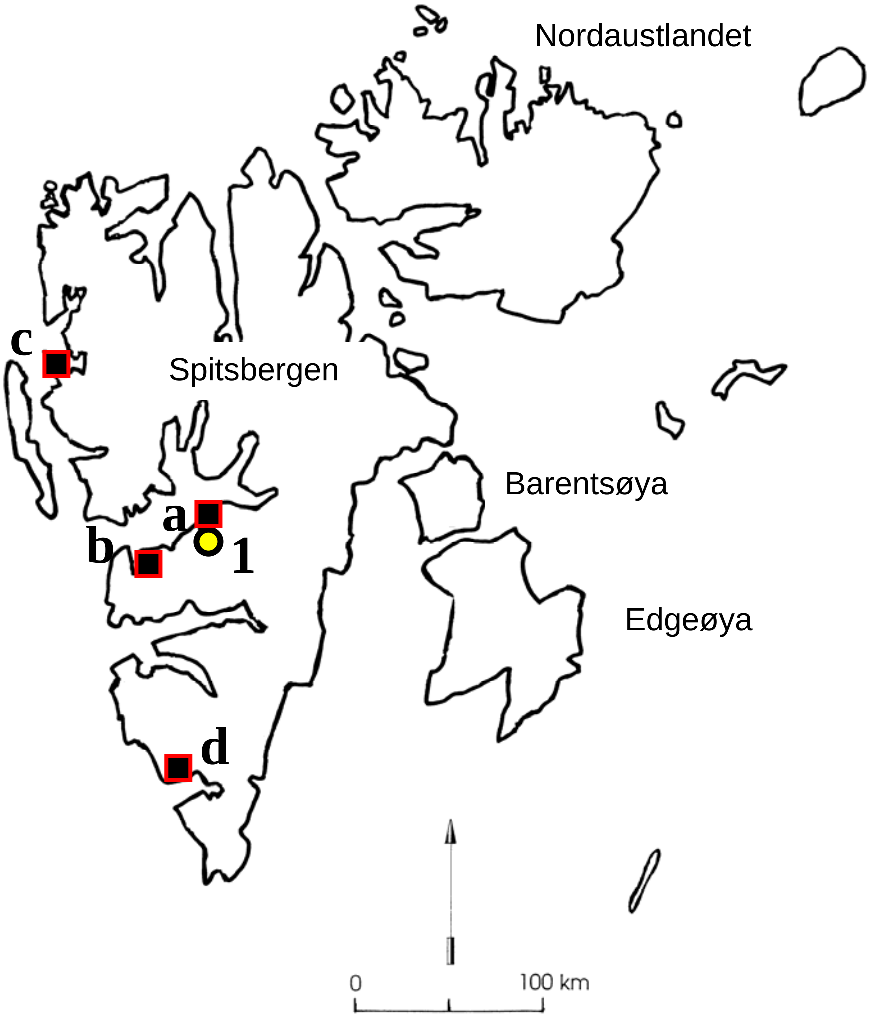

Figure 2.1 The archipelago of Svalbard. The location of the study site is indicated by the circle (**1**). Settlements are indicated by squares: (**a**) Longyearbyen, (**b**) Barentsburg, (**c**) Ny Ålesund and (**d**) Hornsund. Modified from (Hagen, 1993)

2.2 Climatic setting

The climate of Svalbard is strongly affected by the North Atlantic current, which is responsible for the relatively high mean air temperature considering Svalbards northern location. Besides, Svalbard displays unique climate sensitivity (Lamb, 1977; Ahlmann, 1953). At the beginning of the 21st century the mean annual air temperature at sea level in central Spitsbergen is about -5°C (Humlum *et al.*, 2003). Temperature on Svalbard has been recorded since 1912 (fig. 2.2). Most striking for that series is the rapid increase of air temperature until 1920. Since then the fiver year running mean has been relatively stable in

Longyearbyen. Mean monthly temperatures in the warmest month reach 6°C (July), while mean temperatures in the coldest month are -15°C in the period of January to March. Especially in winter, Meteorology in Svalbard is strongly controlled by semi permanent pressure patterns in the northern hemisphere. Two situations are characteristic for Svalbard: 1) If an extension of the Siberian High towards the west occurs, transport of mild air with southerly airflow, from Nordic seas towards Svalbard, causes often heavy snowfall and frequently periods of snowmelt even in the middle of winter. 2) If cold polar air masses extend over Svalbard, the weather is characterized by strong westerly airflow and heavy precipitation. Such inversions in winter lead to periods with calm conditions. In those shallow inversions, warm air overlies the colder air near the terrain surface, leading to a decoupling of surface winds from stronger upper layer winds. These inversions are less frequent in summer and autumn. Low pressure systems (cyclones) moving across Svalbard dominate the weather in those periods (Humlum *et al.*, 2003).

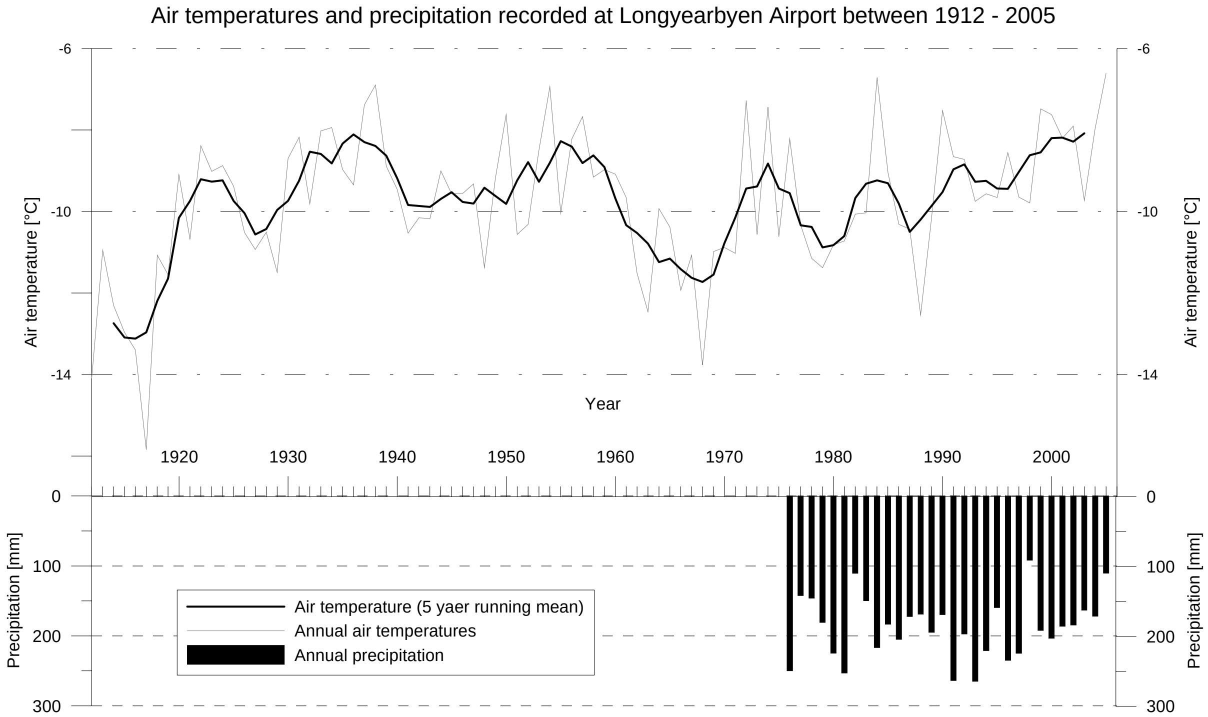

Figure 2.2 Annual mean air temperature and precipitation recorded at Longyearbyen Airport. Data from (eKlima, 2006).

In Svalbard 75 % of the precipitation falls as snow. There is a steep gradient existing where central areas receive a mean annual precipitation of 190 mm while coastal areas receive 3 times as much (Førland and Hanssen-Bauer, 2003; Førland *et al.*, 1997). Anyhow it must be stressed that meteorological stations are few and several problems obtaining reliable measurements still remain. Hansen-Bauer *et al.* (1996) clarifies difficulties for precipitation measurements such as catch deficiency, to distinguish between real and wind-driven snow, as well as topographic aspects. Førland *et al.* (1997) concluded, based on measurements in Ny Ålesund, a correction value of 1.5 in order to obtain the true amount of precipitation. In winter wind speeds on Svalbard are relatively high compared to the summer (fig. 2.3).

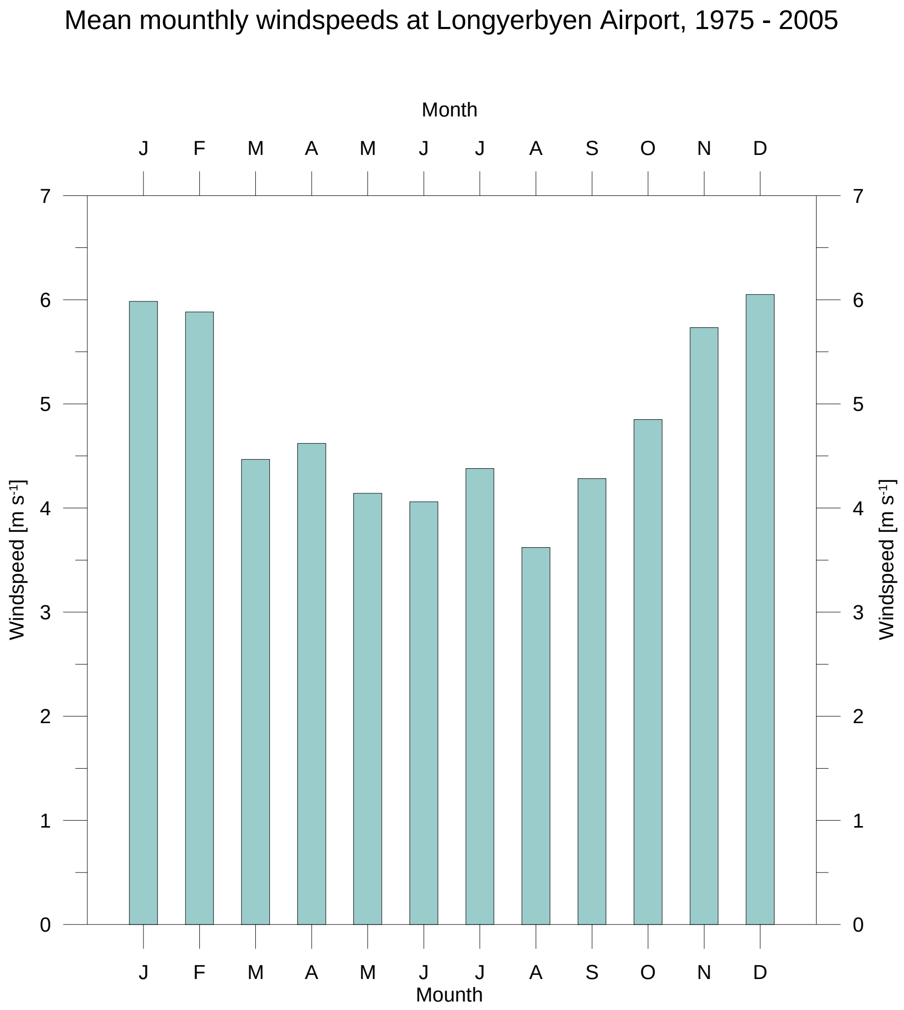

Figure 2.3 Mean monthly wind speeds recorded at Longyearbyen Airport in the period 1975 to 2005. Data obtained from (eKlima, 2006).

2.3 Glaciation on Svalbard, from past to present

The glacial history of Svalbard has been subject to intense research and discussion. During the Pleistocene, Svalbard experienced between 5-8 glaciations extending to the shelf edge during the last 600 kyr BP (Mangerud *et al.*, 1996). After Mangerud *et al.* (1998) there occurred three major glaciations during the early- middle and late Weichselian glaciations interrupted by interstadials. The late Wichselian is by far the most understood and at the same time represents the last glacial maximum (Landvik *et al.*, 1998; Mangerud *et al.*, 1998). Landvik *et al.* (1998) displays evidence for a glaciation which started 25-30 kyr BP ago and lasted until 10-14 kyr BP. Landvik *et al.*, (1998) strongly supports the idea of a Barents Sea ice shelf, confluent with ice over the Kara Sea and extending until the shelf edge. Most of the ice sheet had vanished by 12 kyr BP, but ice remained over Svalbard and adjacent islands and shelf areas. By 10 kyr BP the coasts and major fjords of those islands where ice free.

The following warming period, the Holocene started around 10 kyr BP after the Younger Dryas (11-10 kyr BP) cooling event. During the first half of the Holocene, summer temperatures where higher than today, and many modern disappeared (Snyder *et al.*, 2000; Svendsen *et al.*, 1996). The climatic optimum of the Holocene, with relative warm climate, was followed by a cooling starting at around 5-4 kyr BP, where glaciers advanced and new glaciers formed (Snyder *et al.*, 2000; Mangerud and Svendsen, 1997; Svendsen and Mangerud, 1992). Indications are also found in Nordenskiöldland. There, in the near neighborhood of Bogerbreen, Longyearbreen has advanced from a length of 3 km to 5 km during the last 1100 years. This where shown by dating Relict vegetation found *in situ* under the cold based glacier (Humlum *et al.*, 2005). It is widely accepted that the little Ice Age in the second half of the Holocene coincidences with the Holocene glacial maximum (Svendsen and Mangerud, 1992).

The onset of the Little Ice Age (LIA) on Svalbard was dated, using lake sediments, to the thirteenth or fourteenth century (Mangerud and Svendsen, 1997). This corresponded well with the findings of (Snyder *et al.*, 2000). Both lacustrine and morainic record indicates a two parted LIA divided by a brief warm period (Mangerud and Svendsen, 1997; Werner, 1988). Ice core data suggests the durations of this cold periods to be between AD 1200 and 1500 and between 1700 and 1900 (Gordiyenko *et al.*, 1980). Lefauconnier and Hagen (1990) correlated the measured mass balances with climatic data and reconstructed the mass balance back to 1912. They conclude that the LIA ended at the beginning of the 20th century.

Since then temperature records show a rapid increase in winter temperature from 1912 to 1920 while summer temperatures have been fairly stable (Hagen and Liestøl, 1990). Mean net balances have been negative since then and no significant trend has been observed (Hagen *et al.*, 2003b; Hagen and Liestøl, 1990). At present 60 % of the land is covered by a variety of small- and medium-sized glaciers with a total volume of about 11.000 km³ (Hagen, 1993; Hjelle, 1993). Most dominant are continuous ice masses divided into individual ice streams by mountain ridges and nunataks. Especially in the high alpine mountain regions of west Spitsbergen cirque glaciers are more dominant. Most parts of eastern Spitsbergen and the islands east, such as Edgeøya, Barentsøya and Nordaustlandet, are covered with ice caps. Several of the glaciers terminate in the sea, but are grounded and therefore ice shelves does not exists (Hagen, 1993).

The temperature regime of the glaciers varies but the majority of the glaciers belong to the polythermal type. While the margins and parts of the ablation area are below freezing point, the accumulation area and parts of the ablation area with large ice thicknesses are at the pressure melting point. The majority of the numerous cirque glaciers are of the polar type and the entire ice mass is at a temperature below the pressure melting point (Hagen, 1993).

2.4 Glaciology of Bogerbreen

Bogerbreen was named after Finn Boger, born 1902, who was engaged with the local coal mining company, Store Norske Spitsbergen Kulkompani A/S from 1918 to 1948 (Orvin, 1991). Bogerbreen originates in a single cirque and drains as a valley glacier with a land based terminus. It is a 4.1 km long valley glacier with a surface area of 3.3 km². It ranges from altitudes of 925 m down to 325 m a.s.l. (table 2.1). The glacier itself is draining towards north-west and extends into a well defined terminal moraine located at the drainage divide between Fardalen and Endalen (fig. 2.4). A well defined lateral moraine extends in the north eastern boundary of the lower glacier but is only little pronounced on the north-western side of the glacier. Those terminal and lateral moraines are ice cored and have a debris cover of varying thickness. Theses moraine complexes are mostly decupled from the glacier system (Lukas *et al.*, 2005). It is most likely that the sharp morphological boundary between the moraine and the surroundings mark the postglacial extension during the little ice age (Etzelmüller *et al.*, 1996; Werner, 1988) (fig. 2.5).

Table 2.1 Some selected glaciological key values of Bogerbreen. The values obtained represent the conditions in summer 1990.

Extend	Length along the central flow line [km]	4.1
	Mean width [km]	0.7
	Total surface area [km^2]	3.3
	Blue ice surface area [km^2]	3.2
	Debris covered marginal surface area [km^2]	0.1
	Blue ice / margin area ratio [%]	3
	Maximum depth [m] in 1980. Inferred from (Macheret *et al.*, 1985)	110
Elevation	Maximum [m a.s.l.]	925
	Minimum[m [a.s.l.]	325
	Mean[m [a.s.l.]	
Velocity	Maximum surface velocity in 2004/05 [m yr^{-1}]	1.3
	Minimum surface velocity in 2004/05 [m yr^{-1}]	0.6
Equilibrium Line Altitude 1980. Inferred from (Hagen, 1993)		540

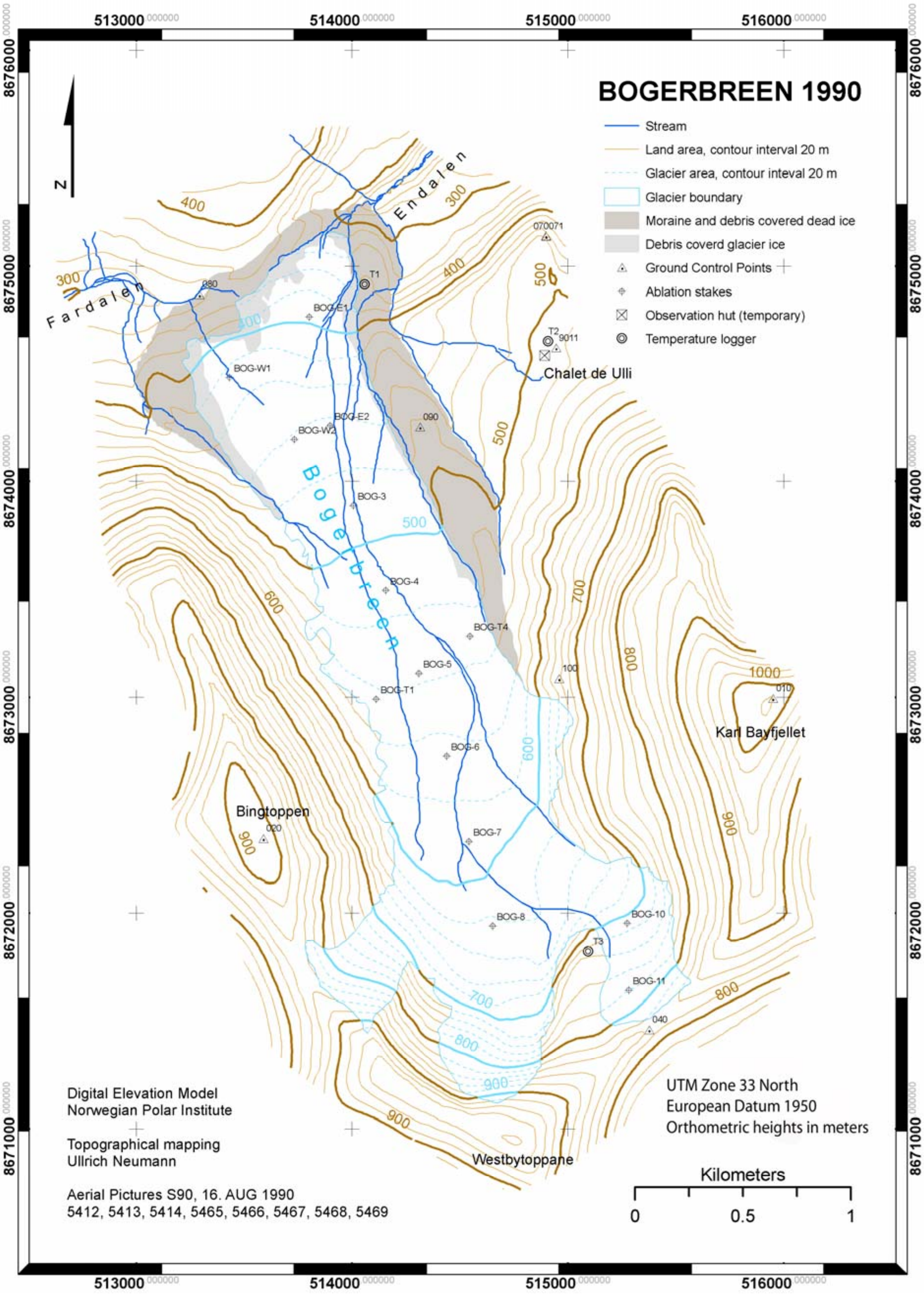

Figure 2.4 Glacier map of Bogerbreen 1990. Orthophotos where used to obtain topographical information. The elevation information's are obtained from Norwegian Polar Institute.

Figure 2.5 Oblique aerial photograph taken 1936. View towards south. Bogerbreen is visible in the center. The glacier terminus is at or close to the position of the terminal moraine marking the postglacial extension during the little ice age.

Most of the glaciers in Svalbard are polythermal. Temperate conditions are mostly met in the accumulation area due to the release of latent heat during refreezing processes and in areas with ice thicknesses greater then 80-100m (Hagen, 1993). Remaining parts are cold based. It is not possible to clearly define the temperature regime of Bogerbreen due to a lack of sufficient investigations. The maximum ice thickness is measured to 110 m by means of radio echo soundings in 1980 (Kotlyakov, 1985). This suggests temperature condition close or at the pressure melting point. Remaining firn areas at the upper cirque could indicate accumulation of temperate ice due to the release of latent heat by refreezing processes. An icing at the terminus appears at many polythermal glaciers (Hagen, 1993). No icings have been observed in the winters between 2003 and 2006. After all, the glacier is most likely cold based with possible limited areas of temperate ice on the upper glacier. Close by glaciers Lars- and Longyearbreen, which have comparable glacier size, show similar thermal regimes (Etzelmüller *et al.*, 2000).

Bogerbreen has a distinct supraglacial drainage pattern with several meltwater channels. Crevasses which could route meltwater into en- or subglacial drainage are few and present only at the upper part of the glacier. Englacial drainage has been observed for shorter distances of a few hundred meters at the lower part of the glacier.

Even though surging glaciers are frequent on Svalbard (Jiskoot *et al.*, 2000; Dowdeswell *et al.*, 1995; Hagen, 1993), Bogerbreen shows no evidence of past surge behavior (Hagen, 1993). The thick debris cover present on Bogerbreen and the absence of push-moraines and ice thrusts are features of many non-surging glaciers in central Spitsbergen (Humlum *et al.*, 2005; Etzelmüller *et al.*, 2000).

3. Methods and data collection

3.1 Collection of primary mass balance parameters

Changes in glacier mass balance can be a result of changes in ablation or accumulation, measuring surface accumulation and ablation is important to understand how glaciers respond to climate variability (Hagen Jon Ove and Reeh, 2004). The direct glaciological method accounts for detailed measurements of summer, winter balance b_s and b_w respectively and calculation of glaciers net mass balance, b_n. The direct glaciological method is considered as the most accurate mass balance measurement method and provides the researcher with a feel for the conditions on site. Despite those valuable facts it is labour intensive and comes with high logistical expenses (Kaser *et al.*, 2003; Patterson, 1994). Even though other methods such as the geodetic method is applied to more and more glaciers due to substantial technological progress, these method presented here seems to be most appropriate for the glacier. The direct glaciological method has been applied to many glaciers on Svalbard and other glaciated parts of the world, in the past and present (Østrem and Brugman, 1991; Gordiyenko *et al.*, 1980). Measuring minimum and maximum balance on a glacier each year is termed the stratigraphic system. It has the advantage of distinguishing between winter- and summer balance which are mostly linked to climatic and topographic parameters. Despite these facts the dates for minimum and maximum values often need to be extrapolated and the stratigraphic record has a time transcressive nature. It has been widely adopted to report mass balances on Svalbard and elsewhere (Østrem and Brugman, 1991; Mayo *et al.*, 1972).

The direct glaciological method has been applied to Bogerbreen to measure surface mass balance within the balance year 2004/05. As described in the previous chapter, winter and summer balances consist of a variety of individual processes contributing to the individual balances. It’s almost impossible to account for all the external and internal processes. Therefore only a selection of most important processes contributing to the glaciers mass balances were chosen to be measured in the field. Winter snow, superimposed ice and avalanche snow were estimated from field data and confined the winter balance.

Surface ablation has been measured to estimate the summer balance. The measured mass balance parameters are reported using the stratigraphic system. Dates for minimum and maximum balances are extrapolated using meteorological observations in Longyearbyen and on site. Field work was carried out between in March 2005 and October 2005. Foregoing preparations started in spring 2004 and the last activities in the field were accomplished in winter 2006. For convenience a observation hut, “Chalet de Ulli”, where placed 1 km east of Bogerbreen at 480 m a.s.l. (fig. 3.1) A total of 42 days of fieldwork were spend on site.

Figure 3.1 The observation hut, "Chalet de Ulli", set up 1 km east of Bogerbreen. The temperature sensor T2 is visible at the pole on the left of the hut. View towards south east during midnight.

3.1.1 Snow sounding measurements

Winter snow accumulation has been estimated with help of snow sounding and transferred to units of water equivalents using a number of snow density profiles. A regular network of sounding was conducted on 30th of April 2005 with 50 to 70 m spacing between the individual sampling points. Converted avalanche probes with 5 cm graduation and length of 320 cm where used for the survey. A total number of 783 soundings where recorded together with the coordinates derived from a handheld Global Positioning System (GPS) unit. The GPS also helped to navigate between the sampling sites. During the collection of the coordinates the horizontal accuracy (X, Y coordinate) was within the range of 20 m as given as a unit from the receiver.

As it can be seen in figure 3.2, the distribution of snow on Bogerbreen is largely heterogeneous. The 783 snow soundings are normally distributed. Snow depth differs between 15 and 320 cm. The mean snow depth is 133 cm with a standard deviation of 30 cm. Largest snow heights exceeding 2 meter, where found on the upper glacier below the headwalls as well as at the western side of the terminus. Less than 50 cm of snow had accumulated along the eastern side at the lower and middle part of the glacier. Raw data, in a digital file format, is given in the appendix.

Due to post accumulation after the initial snow survey, snow probing where repeated along the stake network. The probing revealed post accumulation on most of the stakes. Adjustments to the winter mass balance where made according to the measurements along the stake network (table 3.1). The adjustments consist out of a combination of two independent snow surveys along the stake network (fig 3.3) after the initial winter accumulation has been measured. Snow accumulation measurements where obtained from the lower part of the glacier to stake BOG-4 on May 18th. While from stake BOG-5 on upwards, accumulation measurements where used from a snow survey obtained on June the 4th 2005. The difference of snow level between the initial extensive snow survey and the later measurements were converted into m water equivalent using the mean snow density and the area distribution in each individual elevation band.

Table 3.1 Adjustments added to the initial snow sounding. Thereby each measurement at the stake where assumed to be representative for the individual 50 m elevation band.

Stake	W1	E1	W2	E2	3	4	5
Adjustment [cm]	3	-1	3	-1	2	2	2
Stake	6	7	8	10	11	T1	T2
Adjustment [cm]	4	7	6	1	4	5	4

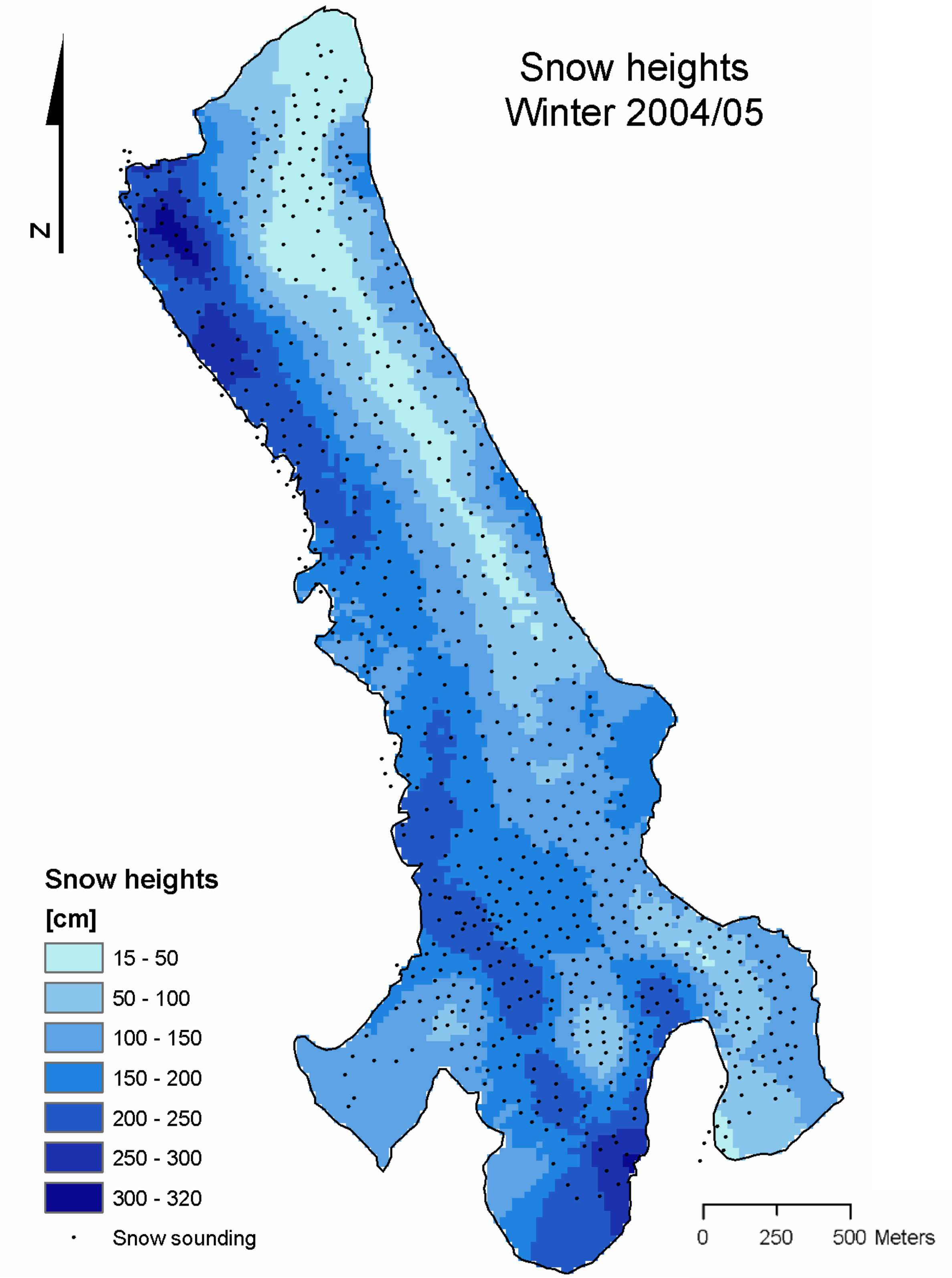

Figure 3.2 Winter snow accumulation on Bogerbreen measured on 30th of April 2005. The black small dots represent the location the snow sounding.

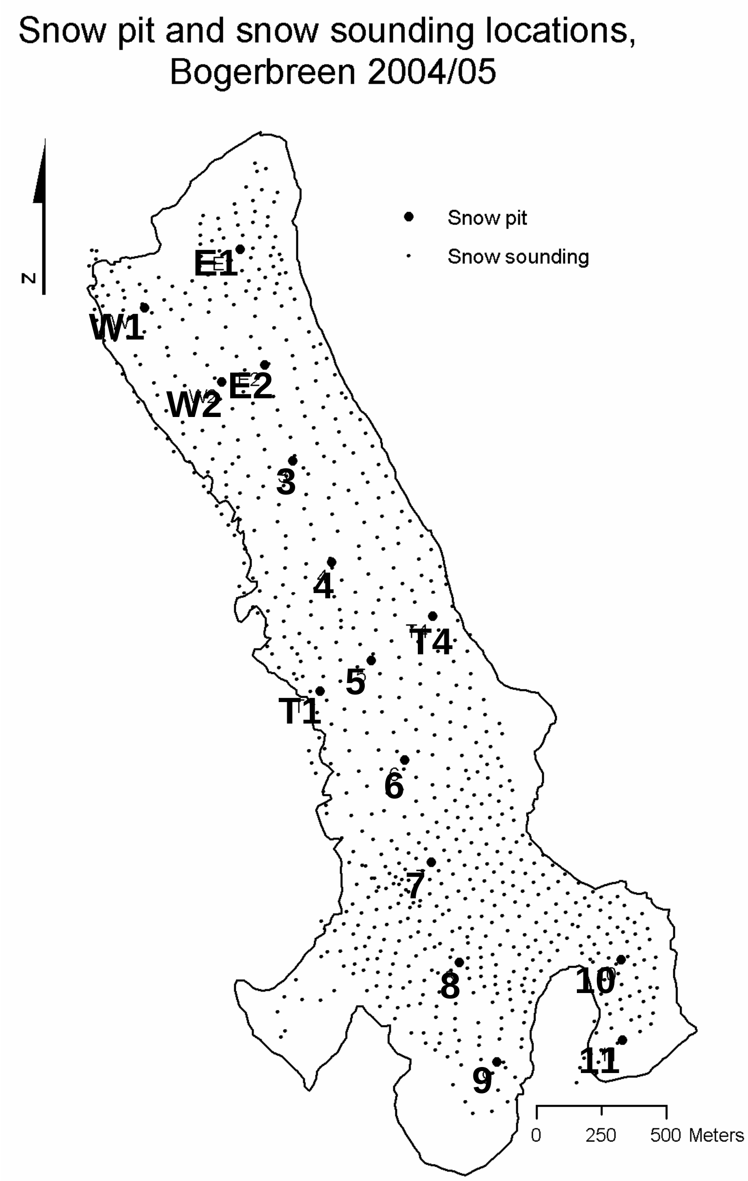

Figure 3.3 Black dots represent stake locations. Stake 9 could not be found after installation but a snowpit where established according to the GPS position.

3.1.2 Snow density measurements

Complementary to the snow soundings, snow densities were measured. This is necessary to be able to convert snow heights into units of volume. A total of 15 snow pits where dug along the stake network (fig. 3.3). The snow pits where established in 3-5 m distance from each stake in order to not disturb the snow stratigraphy for later snow level measurements. On the shaded wall of the snowpit an aluminium cylinder with a length of

19.5 cm and 5.6 cm diameter where used to sample the snow. The cylinder was vertically inserted into the snow for the first 20 cm from the top of the snow pack. The snow filled cylinder was carefully removed from the surrounding snow and the cylinder emptied in a plastic bag. For weighing the sample a PESOLA ® spring balance with 10 g increments was used. The procedure where repeated until the ice surface was meet. Ice layers present in the snow were sampled in the same fashion. The measurements have later been corrected for the mass of the plastic bag, the volume of the sample calculated and the density obtained. The accuracy of this method is difficult to assess. I assumed that accuracy to be the scales increments of about 5 g. Densities where averaged over the individual profiles. An averaged snow density for the entire glacier has been calculated and used to obtain winter accumulation in m water equivalents.

Snow densities as obtained on 30th of April 2005 range from 0.37 g cm^{-3} to 0.47 g cm^{-3} with a mean density of 0.38 g cm^{-3}. The standard deviation has been calculated to be 0.03 g cm^{-3} (fig 3.4). The highest snow density were observed at stake BOG-W1 corresponding to the largest snow accumulation of 230 cm with 0.47 g cm^{-3}. The minimum snow density was found on stake BOG-10. Raw data of snow densities is given in the appendix, in a digital file format.

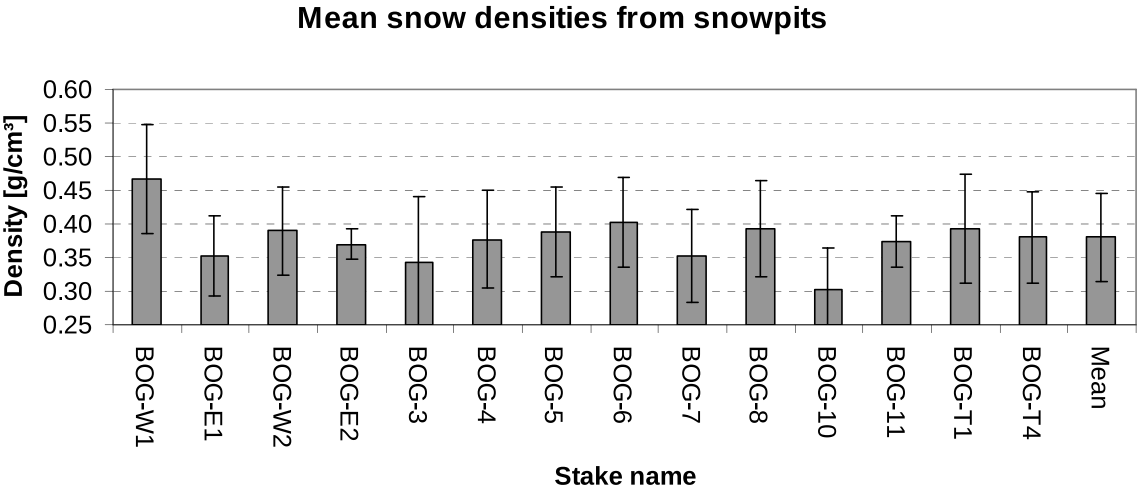

Figure 3.4 Mean snow density at the snow pits along the stake network established 30th of April 2004. Error bars represent one standard deviation of the density variations within the individual snowpit.

3.1.3 Superimposed ice estimation

The accumulation of superimposed ice has been estimated using 2 different methods. There are: drilling of shallow ice cores and the use of artificial reference horizons on the ice surface. Estimation of the autumn accumulation of superimposed ice took place on 30th of April 2005. Shallow ice cores (< 1m) where taken on the stake network (fig. 3.3) with a 7 cm diameter ice drill. The drill equipment consisted of a COWAX ® core barrel and a combustion engine (fig. 3.5. a). A total of 12 cores were examined visually for any prominent change in sediment concentrations, change in the size of captured air bubbles. Thereby it was assumed that superimposed ice is characterized by significant lower sediment content, and partly larger air inclusions then in the remaining lower part of the core. The upper limit of the last years summer surface was assumed to be a distinctive ice layer with an slightly enriched sediment content (fig 3.5. b). Densities of superimposed ice where assumed to be 0.85 g cm^{-3} (Motoyama *et al.*, 2000).

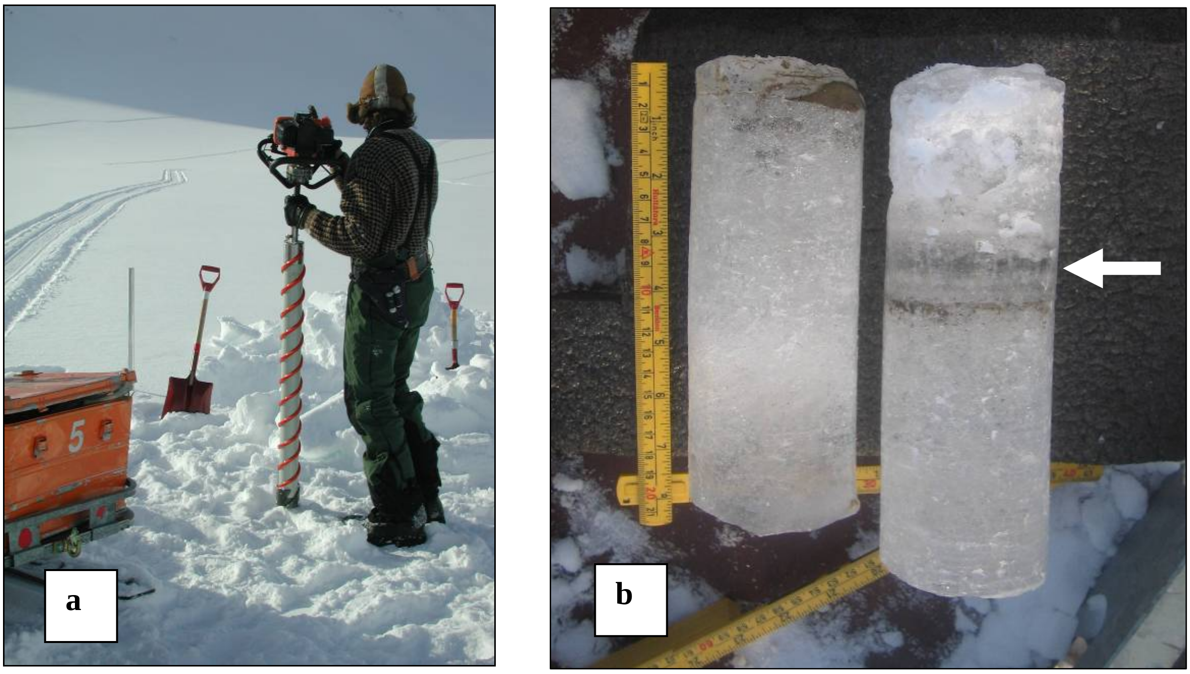

Figure 3.5 Image (**a**) shows the drill equipment used for sampling shallow ice cores. Image (**b**) shows two surface ice cores. On the left core no superimposed ice has formed above the previous year summer surface. The arrow on the right core represents the previous years summer surface above that superimposed ice has formed. Note sediment inclusions, and the bubbly ice on top.

The amount of superimposed ice is very irregular. As figure 3.6 displays, only half the stakes superimposed ice were observed. Autumns superimposed ice reaches thicknesses from 1 to 9 cm while the average thickness is 3 cm. Raw data, of superimposed ice formation is given in the appendix, in a digital file format.

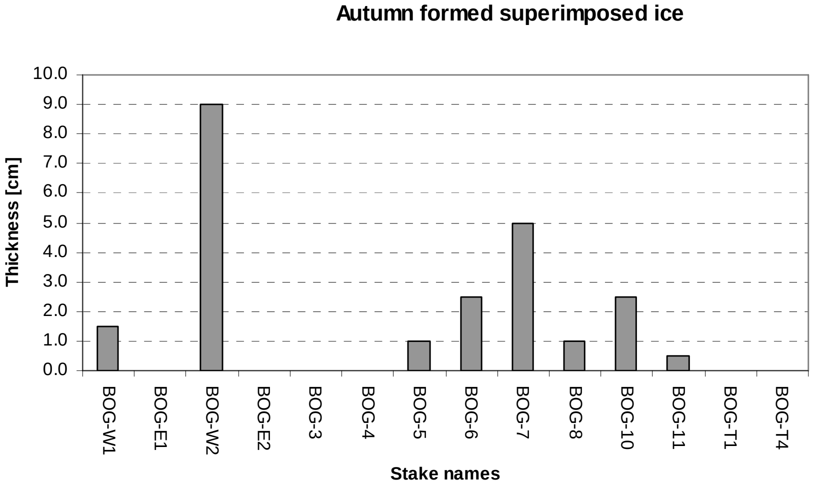

Figure 3.6 Distribution of autumn formed superimposed ice obtained from shallow ice cores.

3.1.4 Avalanche activity observations

The accumulation of avalanche snow has been observed visually on all field visits during field season 2004/05. Observations have been noted down in the notebook with respect of date, location and approximate size.

Small scale loose snow avalanches were observed 30th of April along the north and east facing slops in the upper cirque. Several cornice fall type avalanches were seen on 17th of June with travel distance, from the starting to the runout zone, of 200-300 m and widths up to 15 m. Those avalanches where restricted to the north facing slope in the upper cirque (fig. 3.7). The avalanches were digitised in Esri® ArcGIS, to calculate the area of the deposits. I estimated the influence on the mass balance by assuming a density of 0.55 g cm^{-3} for the deposited snow (McClung and Schaerer, 1993). I applied an estimated mean snow thickness of 1 m in order to calculate the volume of the so derived accumulation.

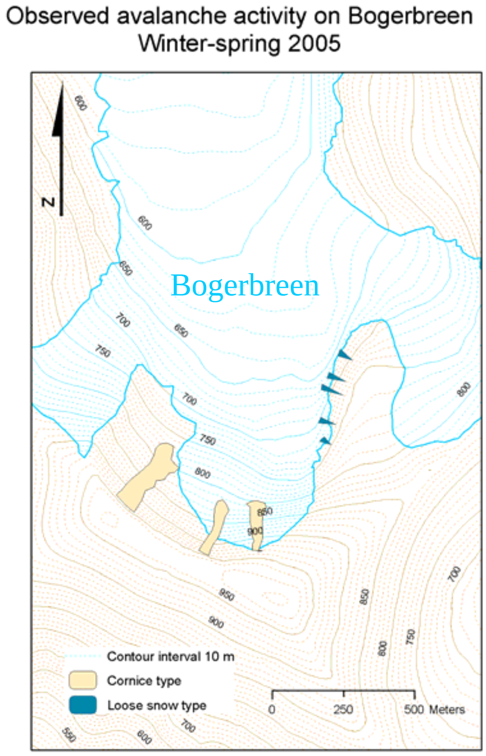

Figure 3.7 Map showing the upper cirque area of Bogerbreen with the observed avalanche activity.

3.1.5 Summer ablation measurements

Following the stratigraphic system, the summer balance is the difference between maximum and minimum balance obtained between two dates (Patterson, 1994; Østrem and Brugman, 1991). Surface ablation has been estimated using the stake method and snow pits. Marker stakes are drilled into the ice surface and used as a reference. The distance between the top of the stake and the ice surface are measured at the end of the winter season and the end of the summer season. The difference between the two is multiplied by the ice density which is considered constant at 900 kg m^{-3}. The result is the balance at that point in m water equivalent. Snowpit studies as described above, give the results for ablation of snow (Kaser *et al.*, 2003; Østrem and Brugman, 1991).

Hagen Jon Ove and Reeh, (2004) and Fountain and Vecchia (1999) concludes that five to ten stakes are usually sufficient for smaller glaciers such as Bogerbreen. A total of 17 stakes were placed on the glacier between 2004 and 2005 (fig. 3.3). 11 stakes were installed in spring 2004 along the centerline (BOG-3 to BOG-9) and a double array at the Terminus (BOG-E1, BOG-W1 BOG-E2 and BOG-W2. During spring 2005, 6 new stakes were drilled to complement the network. A traverse profile was made consisting of 4 stakes (BOG-T1, BOG-T2, BOG-T3 and BOG-T4) and higher altitudes were covered with stake BOG-10 and BOG-11. However stake T2 and T3 where buried during spring 2005 and therefore not included in the mass balance calculations. The aluminium stakes were 5-6 m long and the diameter ranged between 34 mm and 42 mm. The bottom of the stake was sealed with a wooden pluck (fig. 3.8). A 4-5 m deep whole into the glacier was drilled with help of a 5 mm diameter COWAX ® ice drill. Care was taken to drill as vertical as possible. The aluminium rods were then inserted and in some cases, water was added to freeze the stake into the drill hole and reduce settling.

Figure 3.8 Inserting 6m long aluminium stakes into the drill hole. Note the wooden pluck at the bottom of the stake.

The height of the snow has been measured to the previous year’s summer surface. Snow depths were converted into units of volume using the mean density obtained from snow pits as described for winter balance measurements. The minimum balance was observed on October 1st 2005. Since first autumn snow covered the surface of the glacier the summer surface has been measured with help of a probe. In order to enhance accuracy The probe was inserted at four locations around the stake in approximately 30 cm distance and the mean level where reported.

Largest snowmelt has been observed on stake BOG-W1 with 230 cm (fig. 3.9). Little snowmelt on stake BOG-10 and BOG-11 of around 100 cm. Average snowmelt along the stake network was 115 cm. The highest surface lowering of ice recorded was on stake BOG-E1 with 249 cm while the ice surface lowered only 57 cm on stake BOG-7. On average the ice surface lowered 110 cm. While snow ablation is more evenly distributed over the glacier with standard deviations of 45 cm, ice ablation varies more showing standard deviations of 55 cm. Raw data for ablation of ice and snow, are given in the appendix, in a digital file format.

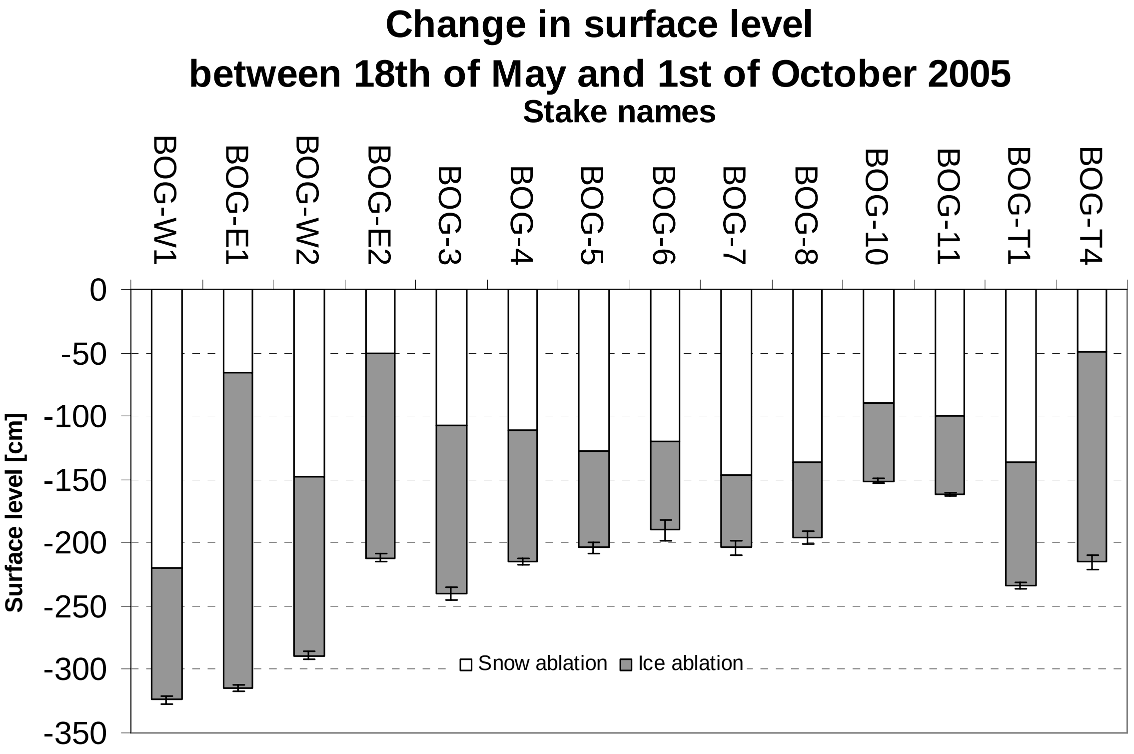

Figure 3.9 Change in surface level observed between 1st of May and 1st of October 2005 along the stake network. Error bars represent one standard deviation.

In order to reduce erroneous levelling measurements, four individual snow depth measurements were taken with the sounding rod on each stake. Each sounding where placed approximately 30 cm away from the stake at 0°, 90°, 180°, 270° relative to north. Standard deviations where then calculated for each stake and individual survey. Results show that on average one standard deviation lays within 4 cm.

In order to estimate the formation of superimposed ice during spring, reference horizons where placed in April 2005 before the spring melt started. Those artificial horizons consisted of 2.5 cm circular plastic tubing with a diameter of 1 meter. The ring was stringed with thin coloured, synthetic, fishing netting (fig. 3.10). Four of the so prepared rings where placed on the ice surface next to stake BOG-7, BOG-8, BOG-10 and BOG-11.

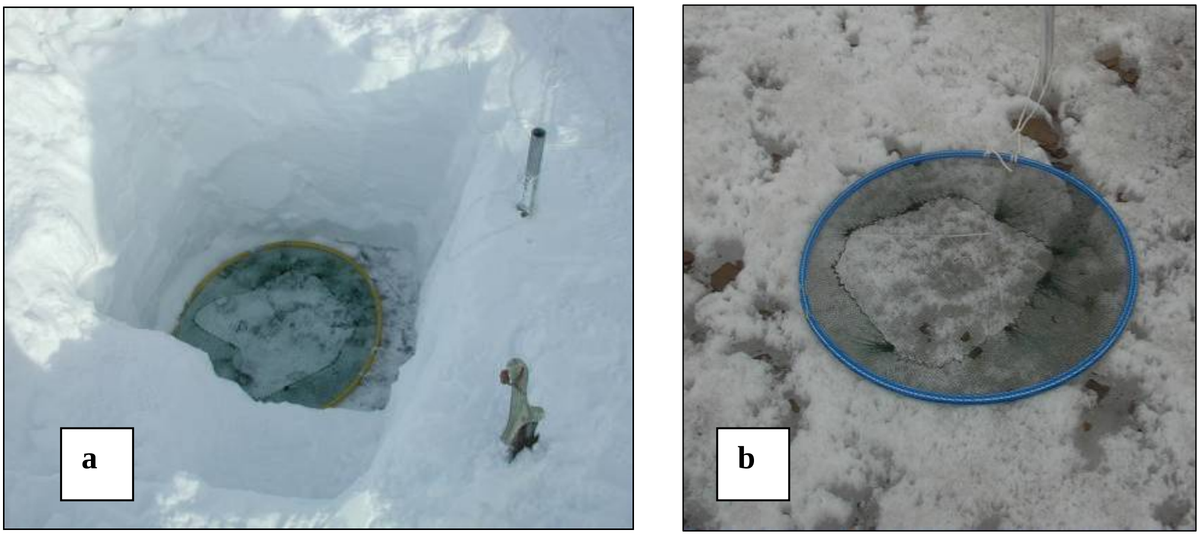

Figure 3.10 Image of the artificial reference horizons used to estimate superimposed ice formed after the winter. Image (**a**) shows the situation during deployment 30th of April 2005 while image (**b**) is taken during the end of the ablation season on 12th of August 2005. No superimposed ice remained at the surface.

Observations on the 17th of August revealed that all artificial horizons remained on the surface. This clearly indicates that no superimposed ice, which might have formed during spring, has remained until then at those four locations (fig 3.10. b). Repeated measurements by levelling along the stake network during summer yielded no significant information concerning timing, distribution and amount of summer formed superimposed ice.

3.1.6 Mass balance calculation

The contour method has been used to calculate Bogerbreens mass balance. There, contours defining areas as equal mass balance are computed or drawn on a map. Then surfaces for areas of equal specific mass balance are calculated within the Geographical Information System (GIS). The product from the specific mass balance and the corresponding areas are then summed over the entire glacier surface to obtain the total change of mass on the glacier. This method is considered as the most accurate to date, it also provides the most detailed information on the spatial variation of mass balance components (Kaser *et al.*, 2003; Østrem and Brugman, 1991).

The specific winter balance consisted of three individual components: Winter snow accumulation, autumn superimposed ice accumulation and adjustments for post snow accumulation after the initial snow survey. Adjustments and superimposed ice where added to the winter snow accumulation measurements correspondingly to every altitude interval. The then combined specific winter balance where determined by interpolation and extrapolation between and outside the point measurements. Interpolation was done in Esri® ArcGIS using a kriging interpolator. The obtained raster had a spatial resolution of 20 m x 20 m. The then computed isolines defined areas of equal specific winter mass balance and areas where calculated within Esri® ArcGIS. Total winter balance B_w where calculated from equation 3.1 in its discrete form,

$$B_w = \sum_{1-n} b_w \Delta S_w \qquad \textbf{(3.1)}$$

where S_w is the area for which the specific winter balance b_w applies.

Determining the total summer balance requires a slightly different approach. Compared to accumulation measurements, ablation measurements are based on only few measuring points. Using computer aided inter- and extrapolation showed rather poor outcome (Kaser *et al.*, 2003; Østrem and Brugman, 1991). Since I was well acquainted with the glacier, I defined areas of equal specific summer balance by hand contouring. Within Esri® ArcGIS, I calculated total summer balance B_s using equation 3.2 in its discrete form,

$$B_s = \sum_{1-n} b_s \Delta S_s \qquad \textbf{(3.2)}$$

where S_s is the area for that the specific summer balance b_s applies. The contour map where then converted into a raster with a spatial resolution of 20m x 20m to make use of the advantages of analysis within GIS.

Instead of calculating total mass balance by hand contouring as described in REF 21, 59. I summed total winter balance and total summer balance within a raster calculation using Esri® ArcGIS. The raster calculation with a resolution of 20m x 20m was based on equation 3.3.

$$b_n = b_w + b_s \qquad \textbf{(3.3)}$$

There, the specific net balance b_n is the sum of the specific winter balance b_w and the specific summer balance b_s for each representing grid cell. Contours where then computed and the total net mass balance B_n where calculated using equation 3.4.

$$B_n = \sum_{1-n} b_n \Delta S_n \qquad \textbf{(3.4)}$$

In equation 3.4 the area S_n with equal specific mass balance b_n are then summed over the entire glacier surface to obtain the total mass balance B_n of the glacier (Kaser *et al.*, 2003).

3.2 Supplemental data

3.2.1 Meteorological observations

Metrological observations are essential to link mass balance to climate (Patterson, 1994). Temperature observations can also give valuable information related to mass balance investigations for example to estimate the onset of the ablation season or local lapse rate. Since the temperature observations on Bogerbreen were very short (4th of June 2005 to 30th of 2006), I correlated the temperatures on Bogerbreen with the long observational record made at Longyearbyen Airport (1912 to present). Wind observations where obtained from a automatic weather station on Gruvefjellet for the balance year 2004 / 05).

The Station at Longyearbyen airport is run by the Norwegian Metrological Institute an located proximately 13 km north of the glacier at an altitude of 23 m a.s.l. Air temperatures were recorded at different locations on Bogerbreen using miniature temperature loggers called Gemini®. The Gemini® Tiny Tag temperature loggers were equipped with a rigid logger housing, a three meter long furled cable and an external sensor. The measuring range lies within -40 to +125 °C, whereas the response time in air is 45 sec and in water 20 sec. The accuracy is ± 0.2 °C (Gemini, 2005). All loggers were positioned on 4th of June 2005 and remained until 30th of June 2006 recording with an interval of 1 hour.

Logger T1 was placed at the terminal moraine at an approximate altitude of 330 m a.s.l. (fig. 3.11. b). The temperature sensor was installed in a 50 cm high, well ventilated stone cairn to prevent direct radiation, snow cover and destructive forces. During repeated visits the stone cairn and the sensor itself remained mostly snow free. Logger T2 was positioned at the plateau east of Bogerbreen at an altitude of 490 m a.s.l. Here the sensor was incorporated into a radiation shield placed 2 m above the ground. The logger housing was damaged during a storm in autumn 2005 and has been replaced by an identical logger (fig. 3.11). Logger T3 was situated at a bedrock outcrop within the cirque at an altitude of 713 m a.s.l.. There, in the same way as for logger T1, a stone cairn was built to protect the logger (see also the map fig. 2.4).

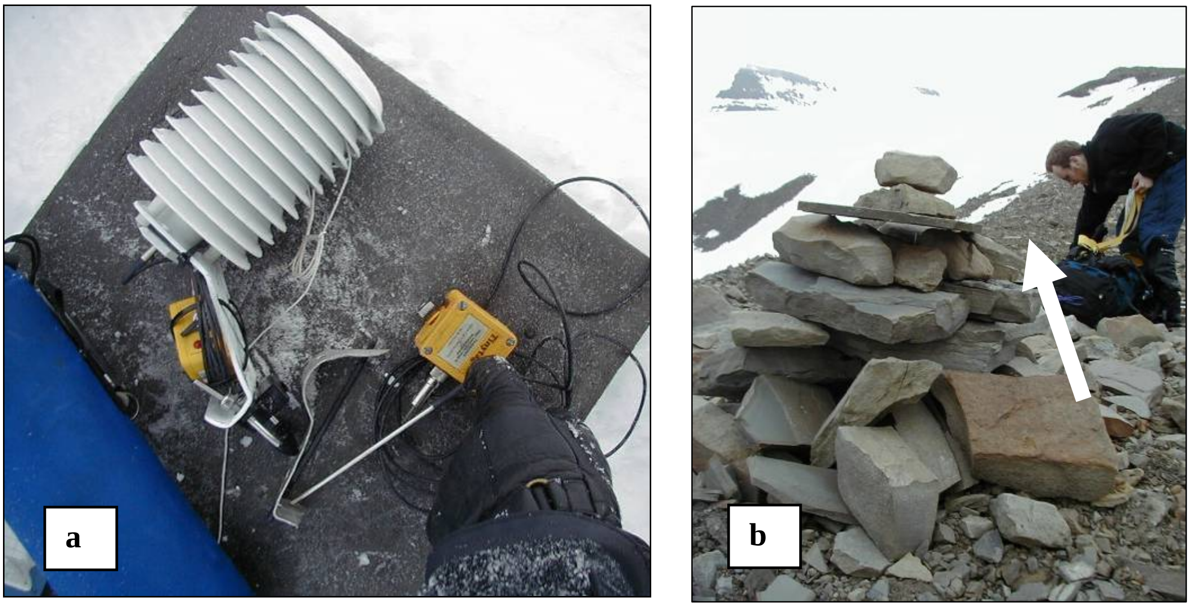

Figure 3.11 Picture (**a**) shows the logger and the sensor T2 (490 m a.s.l.), mounted in a radiation shield. Picture (**b**) is taken on the terminal moraine

showing the stone cairn with logger T1 (330 m a.s.l.) placed in side. The white arrow points at the sensor.

Air Temperatures, at the three stations T1, T2 and T3, were recorded between 5th of June 2005 until 4th of June 2006 (fig. 3.12). The measured air temperatures at T1, T2 and T3 have a rather constant offset towards each other except for a period between start of June and middle July. It is likely that insufficient ventilation inside the stone cairn was the cause of increased temperature. During one of the field visits the logger has been checked and ventilation enhanced. Maximum temperatures where recorded 1st of July 2005 with 16.2 °C at the highest located logger while the minimum temperature were recorded with -24.4 °C at the lowermost logger. Raw data of air temperature recordings, are given in the appendix, in a digital file format.

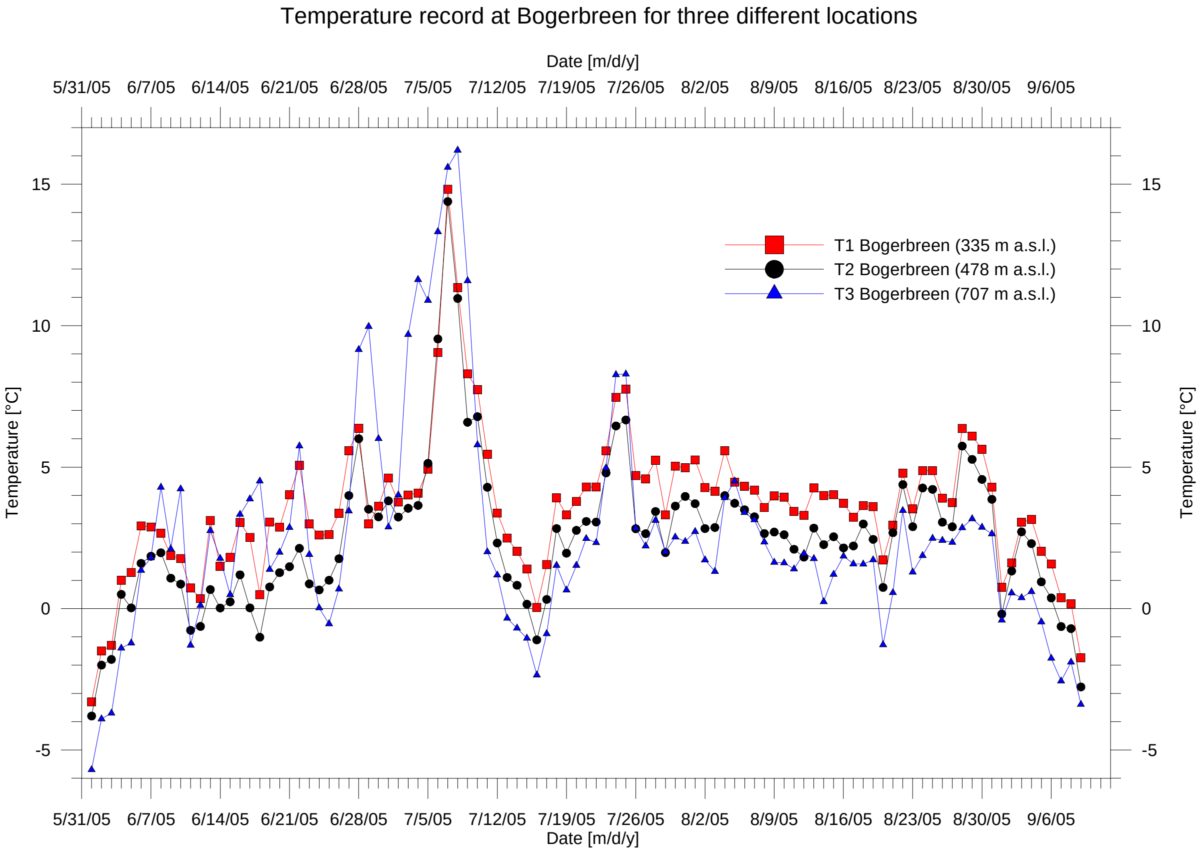

Figure 3.12 Temperature records of the sensors installed at 3 different altitudes on Bogerbreen. Note that the logger T3 shows highest temperatures for the period June to middle July while it is lowest for the rest of the summer season. The sensor being in contact with a rock might be a likely cause of that temperature distribution.

3.2.2 Daily glacier monitoring

Following surface processes on a daily basis can improve our understanding about processes on the glacier. The shifts of snowline during the ablation season and snowfall during summer are measures which can help to better estimate glacier mass balance. With help of a remote digital camera (RDC) such parameter can be analyzed.

A Remote Digital Camera (RDC) has been put up 13th of August 2004. Located at the plateau, south of the Trollstein mountain. The photo direction is towards the south covering the whole Bogerbreen catchment. The RDC is a single unit consisting of a housing, digital camera, controller, battery and a solar panel (fig. 3.13). The camera was programmed to take one picture at 12:00 local solar time. However, during the period between 9th of May to 12th of August 2005 the camera was reprogrammed to take 2 pictures daily. Picture 1 was taken at 02:00 while picture 2 at 19:00 local solar time. The reason was to avoid unwanted reflections from the sun in the camera lens.

Figure 3.13 The remote digital camera was build into a stone cairn at approximately 430 m a.s.l., 1 km north from the terminus of Bogerbreen.

3.2.3 Base map generation

Accurate topographical maps are an important source of information for glaciological investigation (Østrem and Brugman, 1991). In Svalbard topographical maps are available in a scale of 1:100 000. Unfortunately the quality of such maps is only little satisfying if handling smaller surface areas such as Bogerbreen (3.7 km²). Therefore a different approach has been chosen using a digital elevation model within a framework of a Geographical Information System, GIS. The application of such a tool has been previously described by Rippin *et al.* (2003), Etzelmüller and Sulebak, (2000) and Etzelmüller *et al.* (1993). The use of digital elevation models DEM as an important tool for glaciology has been increasing since the last decades (Etzelmueller and Sulebak, 2000; Janson, 1999). The use of DEM's proved to be highly effective for calculating a variety of topographic parameters such as slope and aspect as well for visualizing (Tomlin, 1990). A digital terrain model can be defined as a statistical sampling of the X (longitude), Y (latitude) and Z (elevation) coordinates of the terrain. By interpolation, the elevation can be estimated at any points from its X, Y coordinates. A common way for gathering data for DEM's is by means of photogrammetry (Tomlin, 1990). The purpose of generating an orthophoto in conjunction with the DEM is to produce a topographical map of Bogerbreen. This will be later used for cartographic analysis such as glacier hypsometry and as a base map to display a variety of thematic maps. In orthophotographs distortions due to relief and tilt has been removed (rectified). As a result, the orthophoto possesses the geometrical characteristics of a map with a uniform scale. Orthomaps show a great detail of surface information that is missing from a conventional map (Bannister *et al.*, 1998).

A low resolution (20 m) DEM was available from the Norwegian Polar institute. The DEM is based on aerial survey on 29th of July 1990 with a scale of 1:50 000. The DEM has a 20m resolution. According to Faste pers. comm. (2005), the DEM has a horizontal accuracy (X, Y) of 1 m Root Mean Square Error (RMSE) and 3-5m RMSE in vertical direction (Z coordinates).

An attempt was made to generate a high resolution (5m) DEM using the large scale 1:15 000 aerial pictures of the area obtained by the Norwegian Polar Institute. The aerial pictures of the series S90 where taken on 16th of August 1990 at an approximate flying height of 3000 m.a.s.l. Eight pictures where purchased (S90- 5412, 5413, 5414, 5465, 5466, 5467, 5468, 5469), covering the Bogerbreen catchment. The negatives of the aerial pictures were subsequently digitized with high resolution for use in a digital photogrammetric work station. By solving analytical photogrammetric equations within the workflow, accurate solutions for *X*, *Y* and *Z* coordinates were obtained. Input data consisted of camera parameters, measured photo coordinates and Ground Control Points (GCP). The GCP were surveyed between September 28 and October 1st 2005 with help of differential GPS. The coordinates of the GCP are attached in the appendix.

After the DEM was successfully obtained I generated an orthofoto. Thereby six of the digital image was processed using analytical photogrammetry equations to modify each pixel location according to the scale at that point in the stereo model and the tilt in the photograph. Using this method, it relocates each pixel in the image to a position they would have in a truly vertical photo and with a uniform scale (Wolf and Ghilani, 2002). The orthophotos had been computed via digital image processing using Image Station Base Rectifier Version 8.0 by Z/I IMAGINE® software.

Following licenses were used during the work process: Image Station Digital Mensuration Version 8.0 by Z/I IMAGING ®, Image Station Automatic Elevations Version 8.0 by Z/I IMAGING ®, Image Station Stereo Display Version 8.0 by Z/I IMAGING ®, Micro Station 8.01 by BENTLAY ®.

3.2.4 Surface ice velocity measurements

Even if ice velocities are not required to measure mass balance of a glacier, surface velocities help to define the general character of the glacier. Therefore marker stakes along the central flowline were surveyed in order to estimate ice surface velocities. The method has been common practice since long (Østrem and Brugman, 1991). The relatively new survey method, of using precise global positioning system GPS, proofed to be accurate within cm accuracies by utilizing differential carrier phase measurements (Ayers and Yan, 1995). This has also successfully applied on Kongsvegen, North-West Spitsbergen (Eiken *et al.*, 1997).

A total of 10 maker stakes has been survey using 3 Trimble 5700® GPS receivers in connection with Zephyr Geodetic® antennas. The first survey took place on 14th of April 2004 shortly after marker stakes has been drilled. The survey was repeated roughly a year later in April 12th 2005. The later surveys were conducted in a way to match the previous as close as possible. On both occasions the base receiver was set up at a trigonometric reference point in Longyearbyen “NP 124”. The coordinate of this point where available in World Geodetic Survey 1984 (WGS 84) datum. There the estimated elevation difference between geoidal heights of the WGS 84 datum and orthometric heights is 31.90 m .Two other receivers were alternately placed on top of the marker stakes. The resulting baseline between base receiver and the rovers had a maximum length of 11,4 km. The sampling interval on all receivers was set to 1 second while the individual rovers where placed for a time window of 20 min onto the marker stakes. Coordinates were post processed using Trimble Total Control ™ by Trimble Navigation Limited software. The short occupation periods on the stakes as well as long baseline and obscured sky view in the mountainous contribute further to the mean standard deviation. Therefore it is likely that standard deviations computed from the software underestimates the uncertainty by the factor of three (Trond Eiken, pers. comm.).

3.2.5 Mean geodetic mass balance 1990 to 2004

The geodetic method has been applied on a large number of glaciers and its limitations make it favourable to be applied over larger time steps (Kaser *et al.*, 2003). In order to transform changes of point elevations into mass balance, density must be known and the method need to be applied over the entire glacier in order to full fill the continuity equation (Kaser *et al.*, 2003; Østrem and Brugman, 1991). My intension was to fill the mass balance record-gap, between existing mass balance record from Kotlyakov (1985) and the present data. Therefore I combined the low resolution DEM representing the glacier surface in 1990 with X, Y and Z coordinates obtained from the kinematic differential GPS profiles representative for the summer surface in 2004. Both antenna height and mean snow depth in each elevation band obtained from the snow survey 2005 were subtracted from the Z coordinate of the kinematic DGPS data. The Z coordinate of the 1990 DEM were then extracted and the difference between each point calculated. In areas where no data existed, values where hand extrapolated. I used mean elevation change in each elevation band in conjunction with the hypsometry to obtain the total net mass balance over the period from the balance year 1990/91 to 2003/04.

4. Results

4.1 Primary mass balance results

4.1.1 Winter mass balance

The determination of the winter balance on Bogerbreen showed the following results. The total winter balance b_w were calculated to be 0.55 ± 0.08 m water equivalent (table 4.1). Winter snow contributed most with 97.5 % (0.54 m water equivalent) to the total winter balance.

The spatial variation on Bogerbreen shows little or no longitudinal gradient, while a strong lateral gradient exists (fig. 4.1). There is little or no altitudinal pattern distinguishable, where snow depth increase with increasing altitude. The correlation coefficient between snow depth and elevation is $R^2 = 0.03$ (fig.4.2). The lateral gradient is very dominant at the lower and meddle part of the glacier but less in the higher cirque area. The minimum balances occur both in areas at the terminus and near the headwalls with 0.20 m water equivalent Also along the eastern margin only little accumulation takes place, while along the western margin the accumulation is up to 5 times higher. There maximum accumulation rates up to 1.00 m w.eqv were observed.

Table 4.1 Winter mass balance, both specific and volume at Bogerbreen 2004/05.

Elevation interval [m a.s.l.]	Area [10^3 m ²]	Specific balance [m water equivalent]	Volume balance [10^3 m³]
300 - 350	32.4	0.30	9.8
350 - 400	158.8	0.41	64.6
400 - 450	308.0	0.52	159.5
450 - 500	478.0	0.54	259.1
500 - 550	538.8	0.61	327.1
550 - 600	566.0	0.50	283.6
600 - 650	241.6	0.56	136.3
650 - 700	450.4	0.62	278.8
700 - 750	174.0	0.56	96.6
750 - 800	35.2	0.60	21.1
800 - 850	166.0	0.51	84.0
850 - 900	116.0	0.53	61.8
900 - 950	8.8	0.64	5.6
300 - 950	3274.0	0.55	1787.8

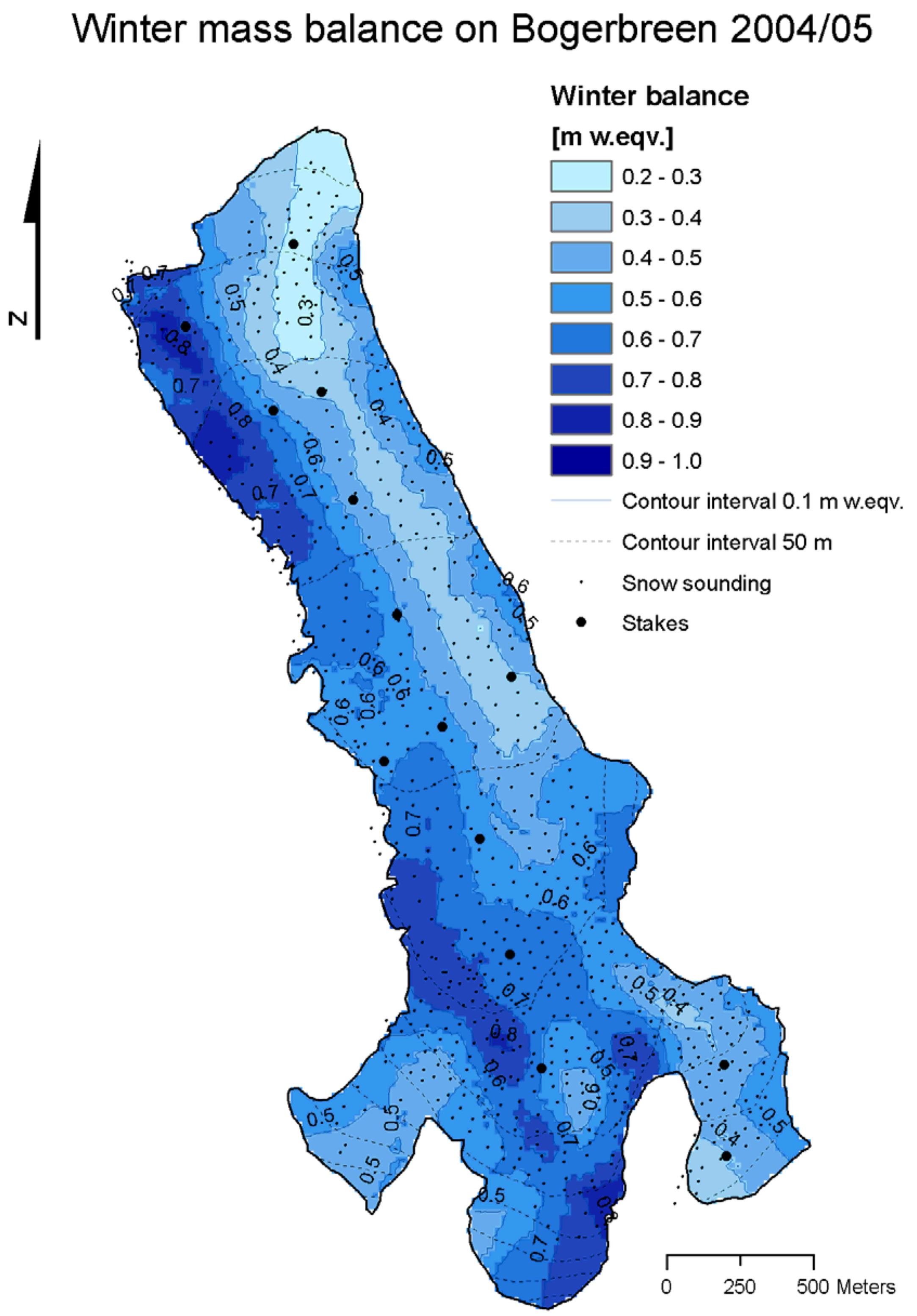

Figure 4.1 The map visualizes the spatial distribution of specific winter balance for the year 2004/05. Small dots represent sounding locations while big dots indicate snow pit locations.

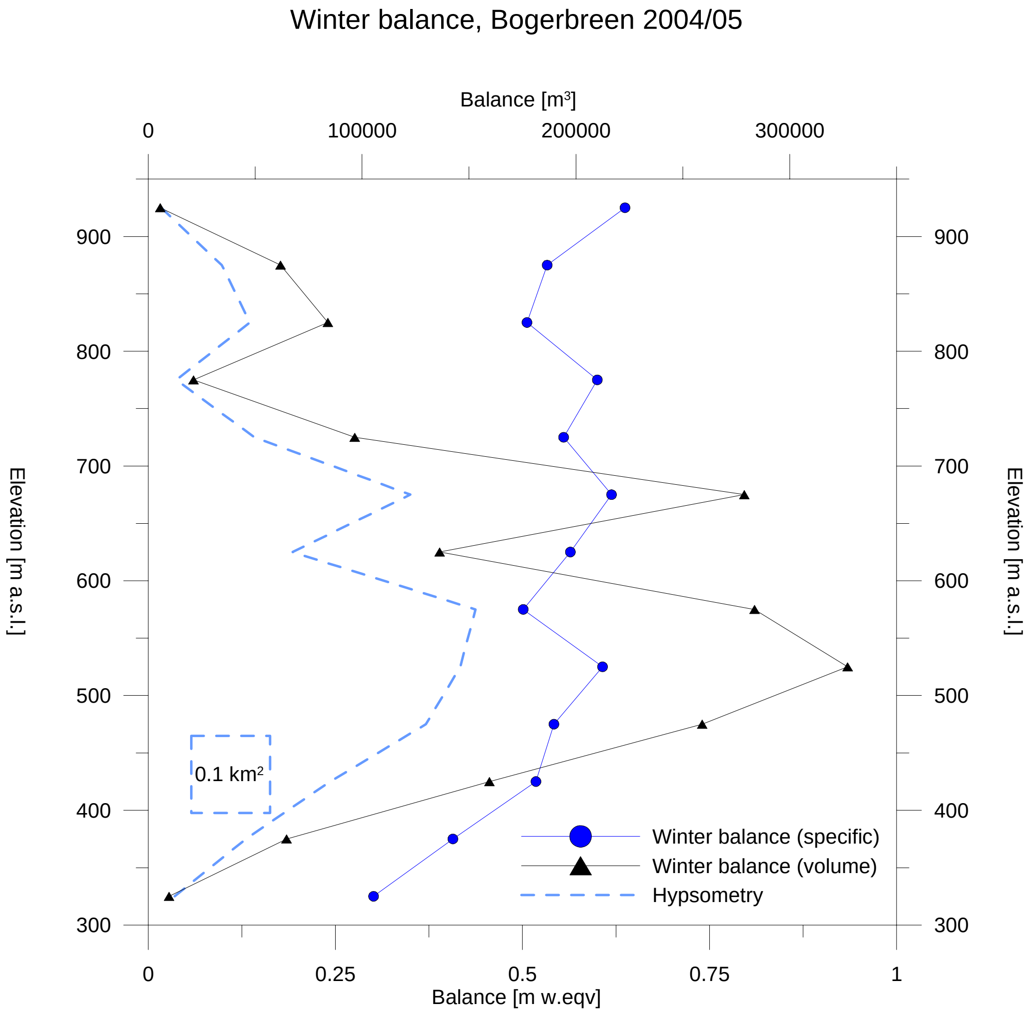

Figure 4.2 Winter mass balance plotted versus altitude for both specific and volume balance.

Superimposed ice has formed only to a little extent, contributing 0.01 m water equivalent, or 2.5 % of the total winter balance. This occurred as autumn superimposed ice (fig. 4.3). Autumn superimposed ice is mainly distributed in the middle, low inclined glacier part (see also fig. 4.19). However the maximum accumulation was found at the 400 – 450 m a.s.l. at the lower part of the glacier but there the surface is more inclined then the middle part.

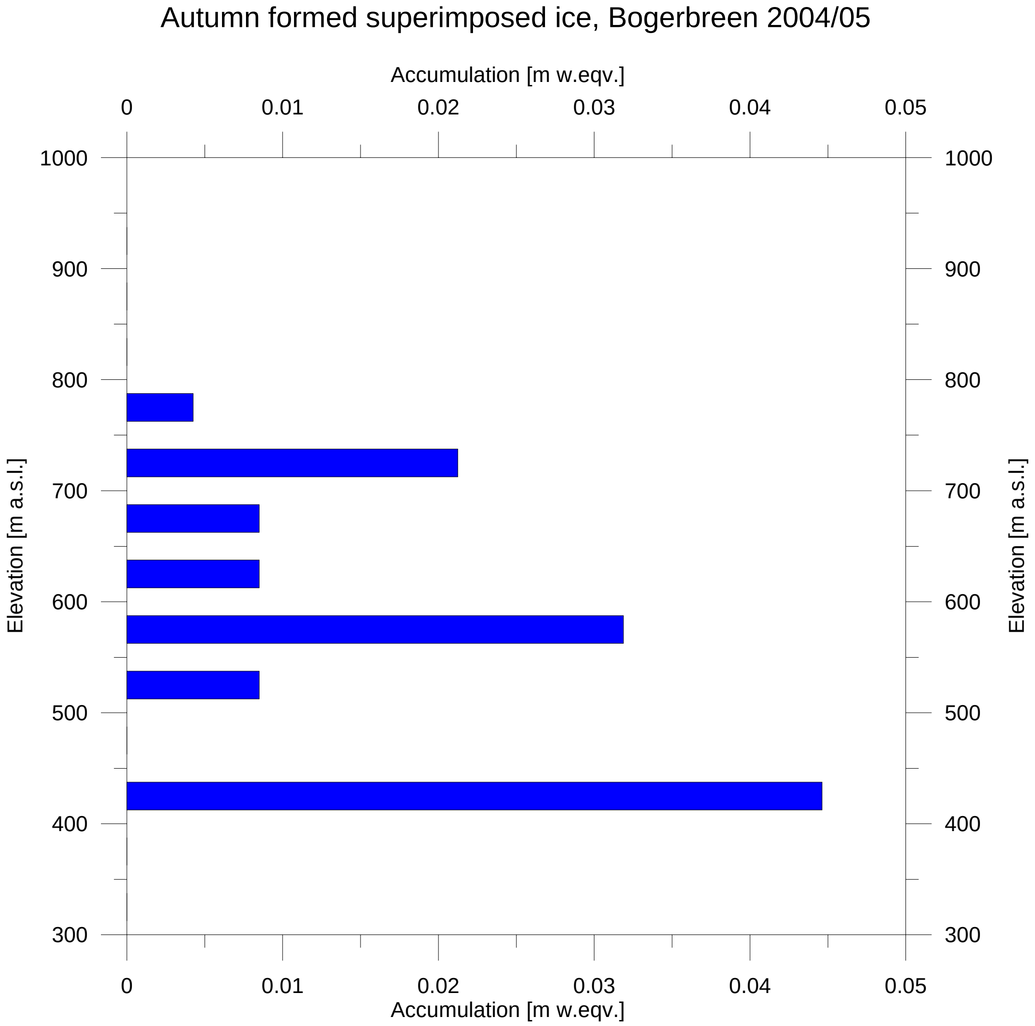

Figure 4.3 Autumn superimposed ice versus elevation as measured by shallow ice cores.

An estimate of the avalanche derived accumulation was calculated to be 0.003 m water equivalent or 0.01 % of the total winter accumulation. It therefore did not contribute significantly to the total winter accumulation

Adjustments of the post winter accumulation after the snow survey on 30th of April contributed 0.0005 m water equivalent or less the 1 % of the winter accumulation. Adjustments are all positive due to late accumulation of snow after the initial snow survey except at the lower elevation band (fig. 4.4). This decrease in snow depth might be caused by compaction of the snow pack, redistribution of snow at the surface, or ablation processes. Generally, it is likely that adjustments are somewhat underestimated due to the fact, that compaction of the snow pack and subsequent increase of snow density was not accounted for.

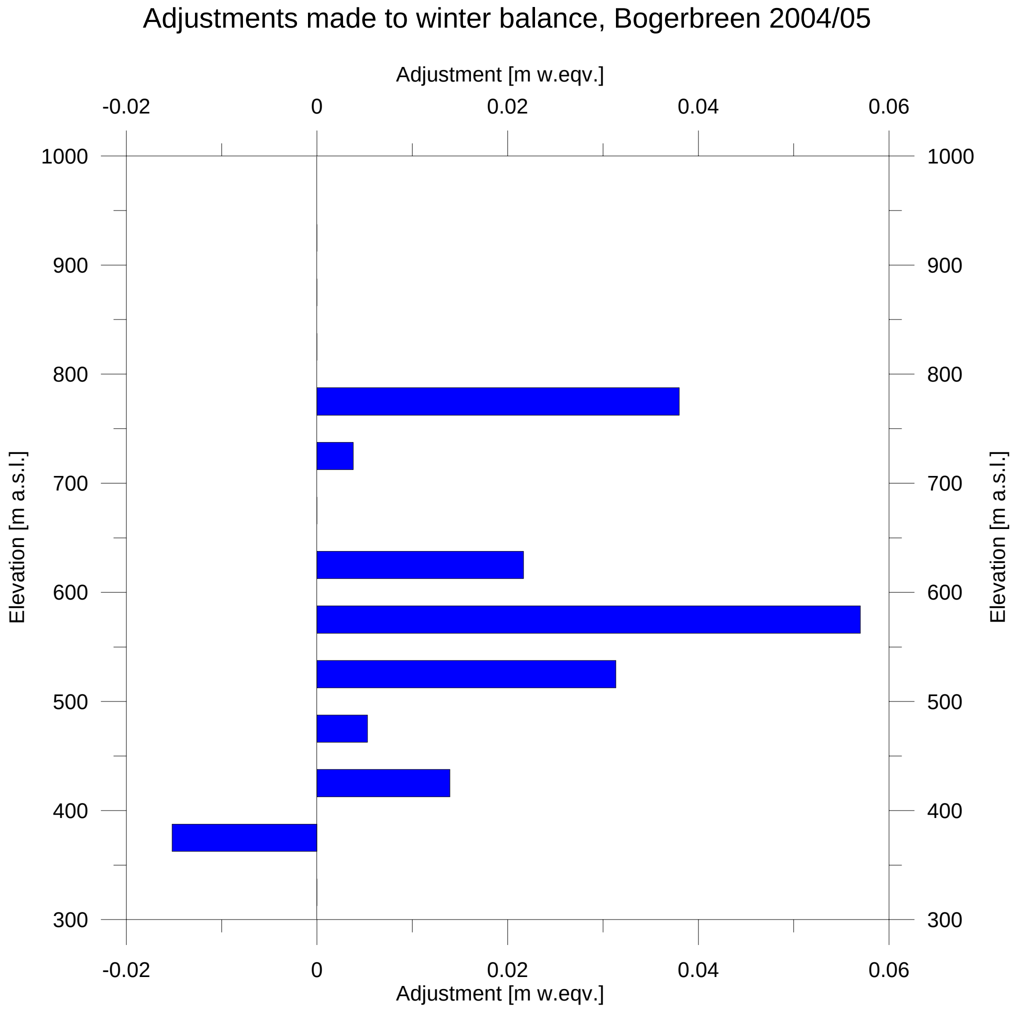

Figure 4.4 The here displayed adjustments were added to the individual elevation intervals to calculate the specific- and total winter balance of Bogerbreen.

4.1.2 Summer mass balance

Bogerbreens summer balance shows a clear trend of decreasing ablation with increase of elevation. The total summer balance was calculated to 1.35 ± 0.24 m water equivalent (table 4.2). Specific summer balances are as high as 2.47 m water equivalent close to the terminus and decrease continuously towards the upper cirque of the glacier (fig. 4.5). There, a minimum is recorded of 0.82 m water equivalent. Towards higher elevations, close to the headwalls, specific balances increase again slightly. Beside the obvious longitudinal gradient a lateral gradient exists, especially in the lower and middle part of the glacier. While isolines of specific summer balance follow elevation contours closely in the upper parts of the glacier, they do not in the lower and middle part of the glacier. The lateral gradient is decreasing with increase in altitude. The enhanced melt patter along the eastern margin for the lower half of the glacier leads to the slight asymmetry of the specific summer balance on Bogerbreen.

The spike like behaviour of the specific summer balance in elevations between 500 and 700 m a.s.l. is due to the influence of the lateral gradient and is also enhanced by the somewhat denser stake network in this elevation interval (fig. 4.6). The volumetric summer balance follows closely the hypsometry. The maximum melt occurs at somewhat lower elevations as the bulk of the glacier area. Summer accumulation occurred on 15th of July 2005. This is evident from the images obtained by the digital camera. The snowfall covered the entire glacier but the accumulation could not be determined in a quantitative manner. No spring superimposed ice remained after the ablation period terminated as observations at the artificial reference horizons showed.

Table 4.2 Summer mass balance, both specific and volume, at Bogerbreen 2004/05

Elevation interval [m a.s.l.]	Area [10^3 m ²]	Specific balance [m water equivalent]	Volume balance [10^3 m³]
300 - 350	32.4	-2.47	-80.0
350 - 400	158.8	-2.41	-382.8
400 - 450	308.0	-2.02	-623.3
450 - 500	478.0	-1.66	-794.9
500 - 550	538.8	-1.12	-601.8
550 - 600	566.0	-1.41	-799.7
600 - 650	241.6	-0.87	-210.1
650 - 700	450.4	-1.02	-457.1
700 - 750	174.0	-0.91	-157.8
750 - 800	35.2	-0.93	-32.7
800 - 850	166.0	-0.92	-151.8
850 - 900	116.0	-0.93	-107.8
900 - 950	8.8	-0.93	-8.1
300 - 950	3274.0	-1.35	-4408.6

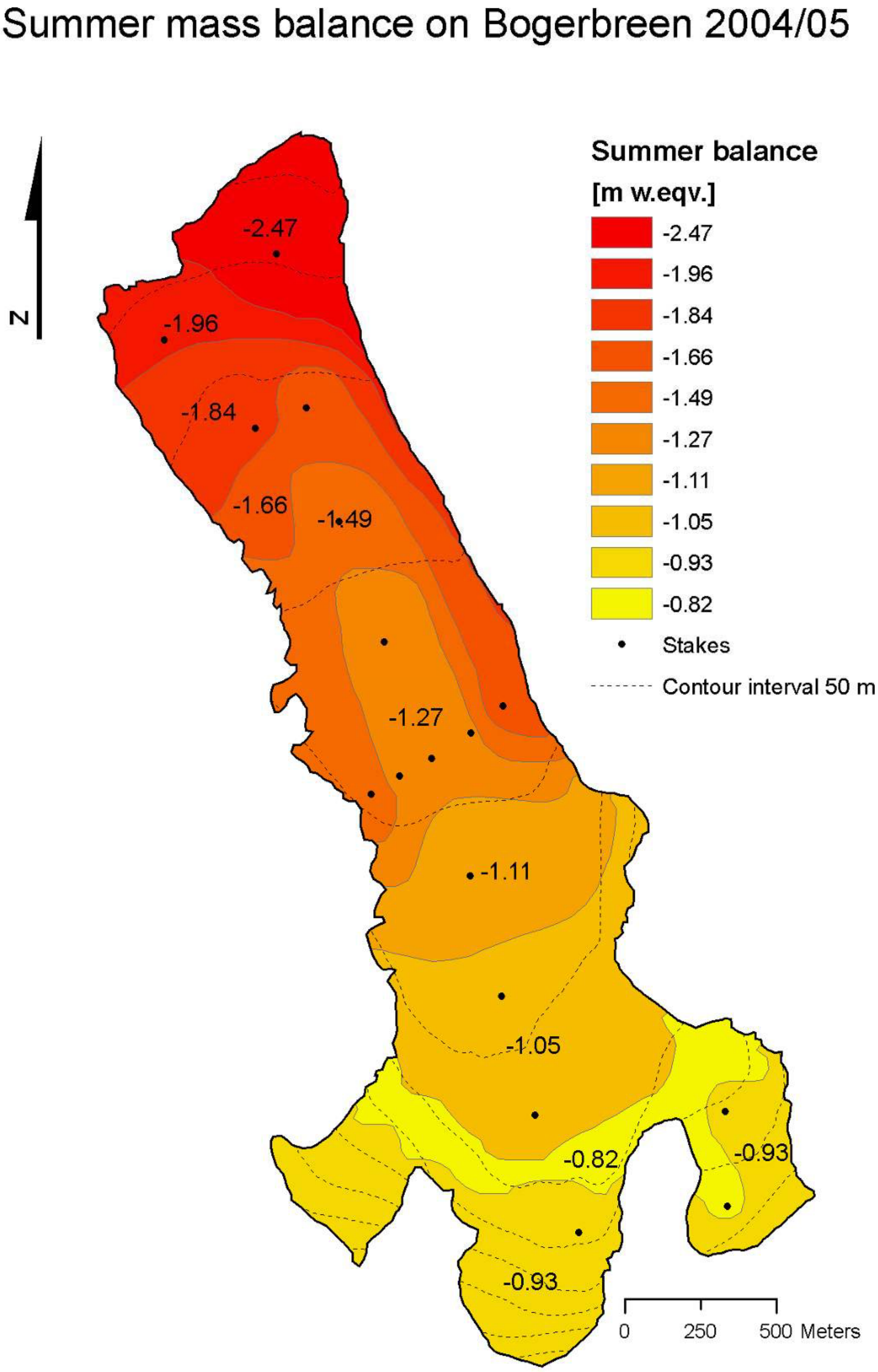

Figure 4.5 Map showing the spatial distribution of specific summer mass balance 2004/05. Stake locations are marked with black dots.

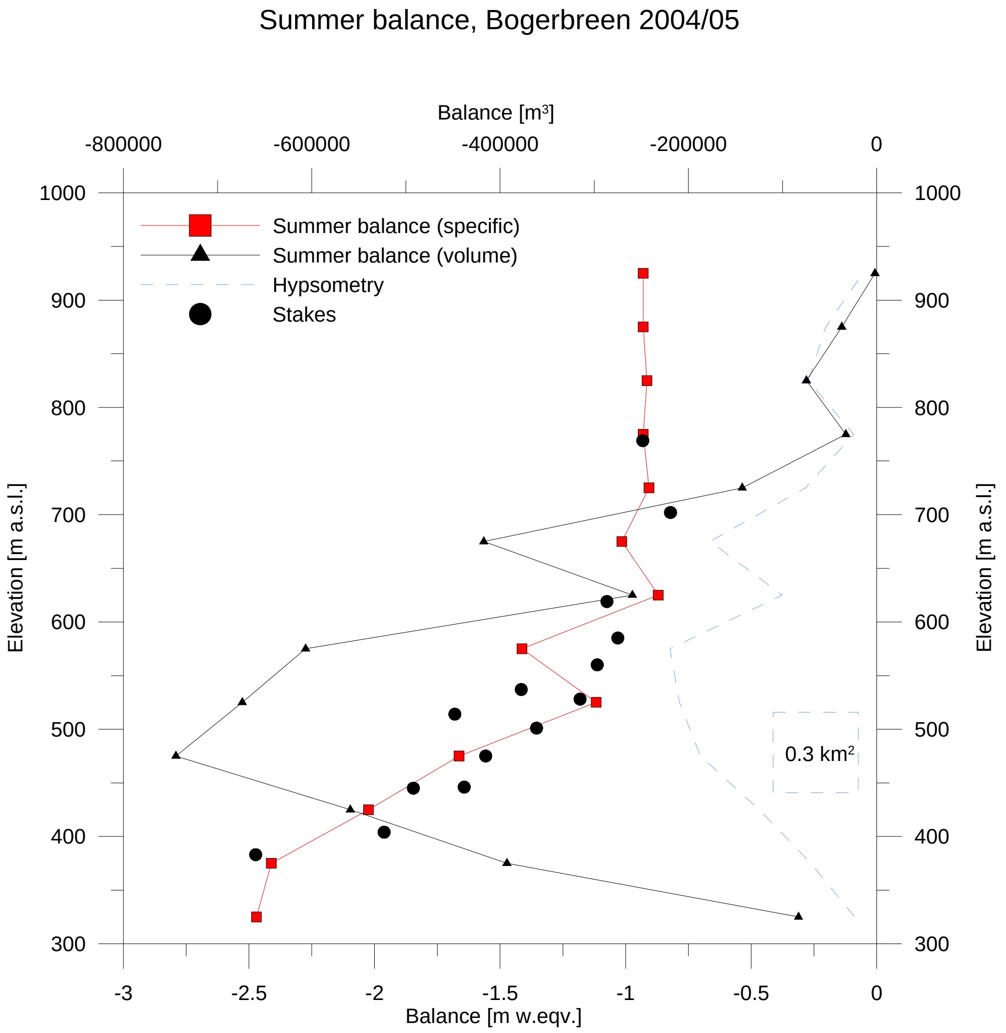

Figure 4.6 Plot of summer balance versus elevation for both, specific and volume balance 2004/05.

4.1.3 Net mass balance

The calculated net balance is the result from the sum of the winter and the summer balance. The total net mass balance were calculated to -0.80 ± 0.20m water equivalent (table 4.3). Specific net balances were negative over the entire glacier surface (fig.4.7). A strong gradient of mass balance exists versus elevation. Lowest specific mass balances occurred at the north eastern terminus with -2.25 m water equivalent while highest net balances of 0.20 m water equivalent are present close to the headwalls in the upper part of the glacier. The longitudinal gradient is largest at the lower half of the glacier while it becomes almost cero at the upper half of the glacier above 625 m elevation (fig.4.8). At the same time I find also a dominant lateral gradient which is dominating the lower half of the glacier. A table with mean specific net balance is given below (table 4.3). The maximum volume balance occurs at the elevation interval of 450 – 500 m a.s.l. and 550 – 600 m a.s.l. and the occurrence is identical with the bulk of the glacier area (fig. 4.8).

Table 4.1 Net mass balance, both specific and volume, at Bogerbreen 2004/05

Elevation interval [m a.s.l.]	Area [10^3 m ²]	Specific balance [m water equivalent]	Volume balance [10^3 m³]
300 - 350	32.4	-2.17	-70.2
350 - 400	158.8	-2.00	-318.2
400 - 450	308.0	-1.50	-462.9
450 - 500	478.0	-1.12	-536.3
500 - 550	538.8	-0.51	-274.8
550 - 600	566.0	-0.91	-515.6
600 - 650	241.6	-0.31	-74.2
650 - 700	450.4	-0.40	-178.8
700 - 750	174.0	-0.35	-61.1
750 - 800	35.2	-0.33	-11.6
800 - 850	166.0	-0.41	-67.9
850 - 900	116.0	-0.40	-46.1
900 - 950	8.8	-0.30	-2.6
300 - 950	3274.0	-0.80	-2620.3

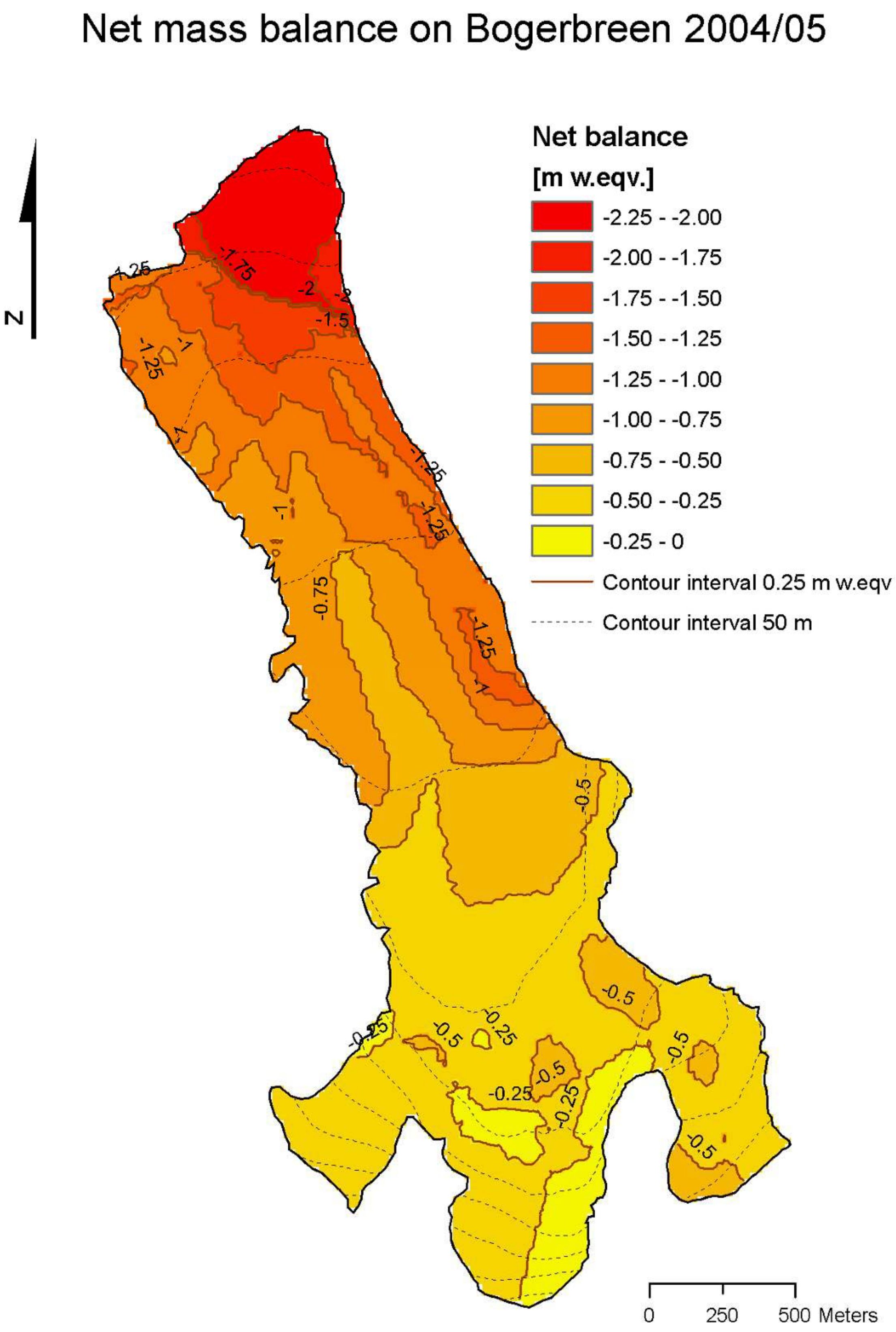

Figure 4.7 Map of the spatial variation of specific net balance at Bogerbreen 2004/05.

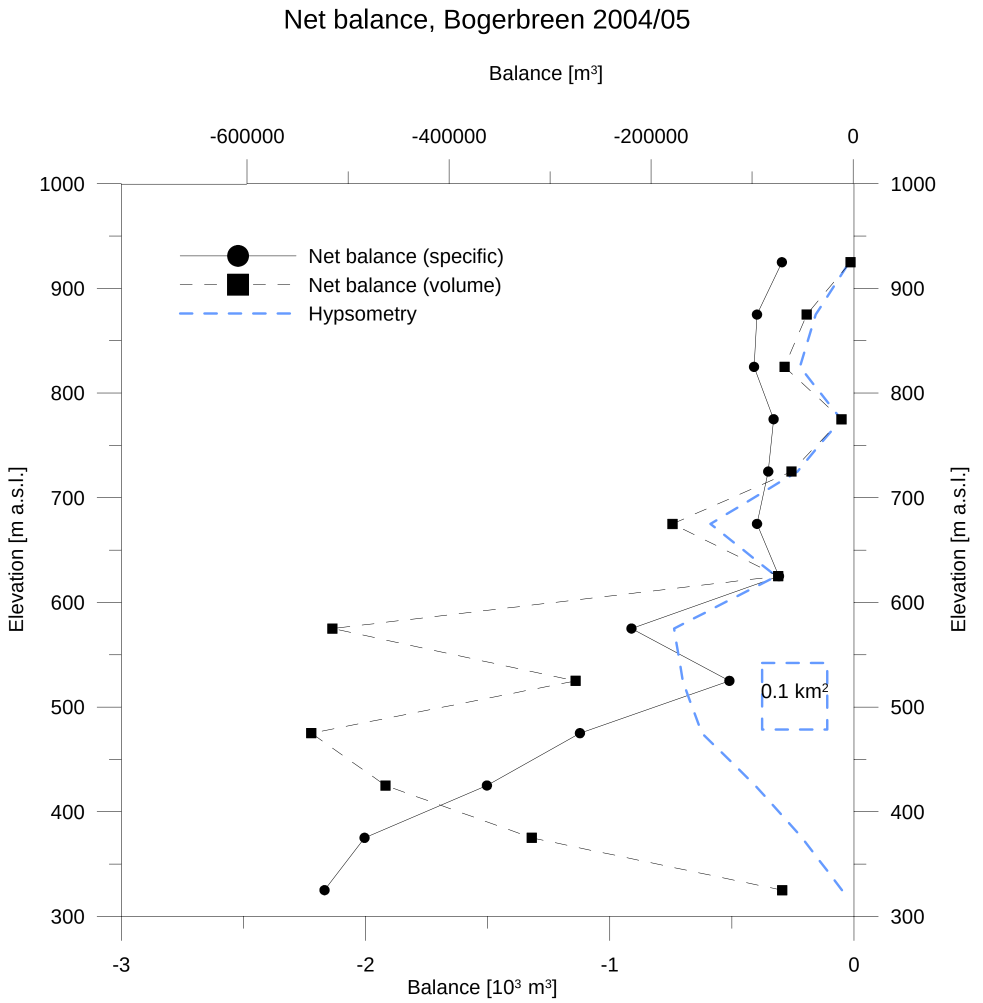

Figure 4.8 Net mass balance, both specific and volume at Bogerbreen 2004/05

4.2 Error analysis

Even though a large effort was made to measure winter- and summer balance components as accurately as possible the direct glaciological method has its limitations. In general systematic and random errors are difficult to determine accurately for traditional mass balance measurements (Østrem and Haakensen, 1999; Janson, 1999; Østrem and Brugman, 1991).

4.2.1 Winter mass balance

Systematic errors in measuring winter balance can arise by misinterpretation of the summer surface during probing (Janson, 1999). On Bogerbreen a clear summer surface was met in the upper and middle part of the glacier due to the absence of firn. However ice lenses up to 1.5 cm thickness were found present in the snow pack. Those are derived by several rain and melt events during accumulation period. Even though control was obtained by six snow pits at the lower part of the glacier it can not be clearly excluded that probing was conducted at a false summer surface. An attempt was made to calculate the effect on such error. For the elevations from 350 m – 500 m a.s.l. where ice lenses were present the error constituting from 5 % and 10 % of the accumulation were 0.03 m water equivalent and 0.05 m water equivalent respectively. This error corresponds to underestimation of the winter balance between 5 % and 10 %. This makes it clear that care must be taken when probing in areas with ice lenses present in the snow pack. I made tests using a 12 mm steel probe. Ice lenses above 2 to 2.5 cm thickness could not be penetrated without considerable force and ice lenses exceeding 3 cm where not penetrated at all. Another factor is the uncertainty for the probe remaining vertical in the snow column or not. But since snow depth is rather little on Bogerbreen this error is neglected here.

Random error is introduced while probing due to miss readings for example. Janson (1999) states that the error increases in importance with decreasing probing points on the glacier. On Bogerbreen 783 probing points where used to estimate winter balance compared to 300 points measured on Storglacären, Sweden for example. This gives me the advantage to spot big erroneous probing easily just by visual inspection of the interpolated snow map. Artefacts can be located and removed from the data sheet. After Janson (1999) smaller errors average out.

Another factor for uncertainty is introduced through the density measurements and the following conversion from snow depth into meter of water equivalents. I estimated the precision of the scale used to measure snow depth to be within 5 g. The effect on the winter balance is ± 0.08 m water equivalent corresponding to 15 % of the total winter balance.

Janson (1999) compared several methods of converting snow depth into units of volume using density measurements from a number of snow pits. He distinguished beside the average density applied over the entire glacier, also the Thyssen polygon approach where a specific density is representative for a certain area and a third approach where density is representative for snow with similar depth. I used a simple approach as a reference value and used the mean density obtained from 15 snow pits for the conversion. Comparison showed that by using average density, area density and depth density approach, the total winter balance were 0.55 m water equivalent, 0.53 m water equivalent and 0.55 m water equivalent respectively. As shown, the winter balance seems to be rather insensitive to the method applied.

Another source of uncertainty arises by interpolation or extrapolation in areas that are not probed. On Bogerbreen 9 % of the total area remained unprobed mainly consisting of steep headwalls and avalanche prone slopes. As mentioned by Janson (1999) snow depth in this areas is generally very heterogeneous. This were proved also by observations on Bogerbreen where snow accumulation were largest distal to the head wall and decreased towards the headwall where bare ice or rocks where visible. If we assume that only 50 % of the accumulation is deposited in such unprobed areas the resulting error is ±0.02 m water equivalent representing 4 % of the total winter balance.

Uncertainties appear also in areas where snow probing might not be representative for the snow depth. Janson (1999) argues that in marginal areas where snow depth varies significantly over short distances, probing might underestimate the snow depth and add to the uncertainties. However on Bogerbreen snow probing was performed in a denser network than on Storglacären minimizing the effect. Still, unprobed areas at the lower part of the glacier summed up to be 2 % of the total area. As calculations show, even thought if we double the accumulation in this unprobed areas, it would not affect the winter balance by more the 0.01 m water equivalent representing 2 % of the total winter balance.

Since superimposed ice and adjustments from new snow survey add to the total winter balance, I considered there uncertainties. Both superimposed ice and adjustments contribute only with 0.01 m water equivalent and 0.02 m water equivalent respectively to the total winter balance. Even though superimposed ice varies significantly spatially (Obleitner and Lehning, 2004; Motoyama *et al.*, 2000), it is unlikely that the uncertainties in these parameters contribute significantly to the total winter balance.

Computer aided geostatistical interpolation of winter accumulation has the advantage being little time consuming and reduced random uncertainties compared to manually contouring of winter balance (Hock and Jensen, 1999; Østrem and Brugman, 1991).

Nevertheless the method has its limitations. Hock and Jensen (1999) describes the effect of various interpolation parameters using kriging interpolation. I calculated winter balance using different kriging parameters for the semivariogram model used within ESRI ArcGIS ®. Results on Bogerbreen show only insignificant variations of the spatial mean and total winter mass balance did not change significantly. To further evaluate the quality of the obtained winter balance I used an elevation interval approach. I used mean snow depth obtained by probing for each 50 m elevation interval. I linearly extrapolated into unprobed areas. The result of 0.53 m w.eqv deviated from the contouring method by + 4 % of the total winter balance (fig. 4.9). Again the high probing density seems to very well represent winter balance also under simplified calculation. It is very likely that the deviation from the reference value is only due to differences into the extrapolated areas by using either computer aided or manual interpolation. I assume that that kriging slightly overestimates the true winter balance due to overestimation of snow depth in the upper cirque of the glacier.

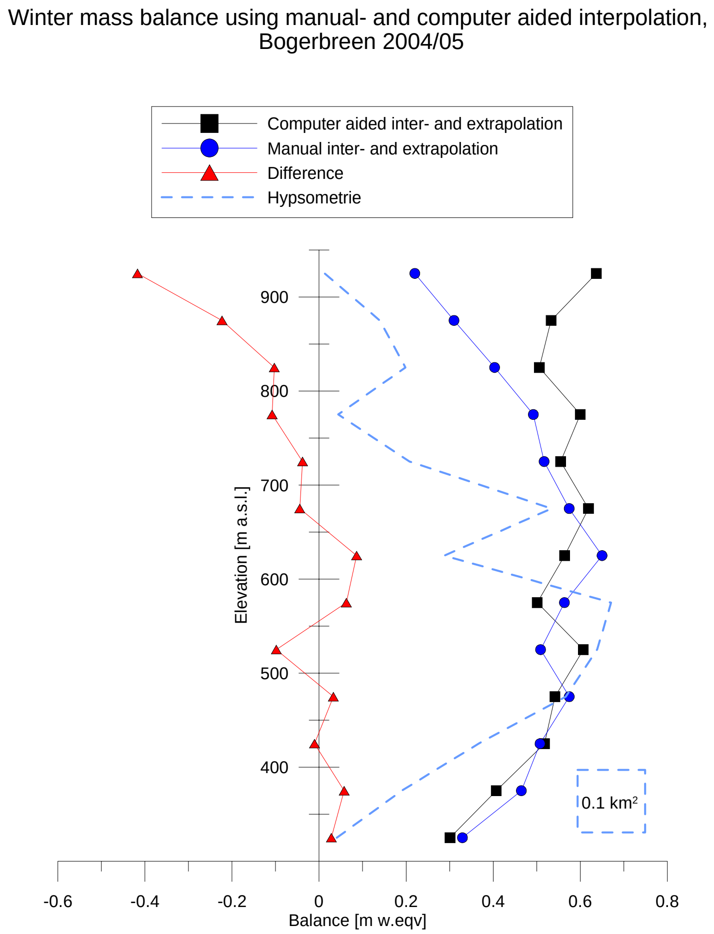

Figure 4.9 Plot showing the spatial difference of specific winter balances in each elevation interval for hand and computer aided calculation of the winter balance.

To summarize the analysis about winter balance uncertainties, I therefore argue that my derived estimate for the total winter balance on Bogerbreen of 0.55 m water equivalent is accurate within ± 0.08 m water equivalent or 15 % of the total winter balance. The winter mass balance is most sensitive to uncertainties in density determination. Errors up to ± 0.08 m water equivalent corresponding to 15 % of the total winter balance can easily occur if density measurements are not made with caution. Errors up to 10 % can occur if a false summer surface is probed (ice lenses). Other above mentioned uncertainties affect the total winter balance within only with 0.03 m water equivalent or 5 % of the total summer balance.

4.2.2 Summer balance

An effort was made to evaluate the quality of the hand contouring in three steps. The area distribution of the 10 defined ablation areas was altered with a Gaussian random number. The Gaussian random numbers had the same variance as the original area distribution. This approach gives a feel for both sensitivity of total summer balance against misjudging contouring procedure as well as errors introduced within GIS software by determining correct areas. The 30 random numbers where then used to individually calculate the total summer balance. The so collected summer balance values range between ± 0.12 m water equivalent representing 9 % of the total summer balance (fig. 4.10).

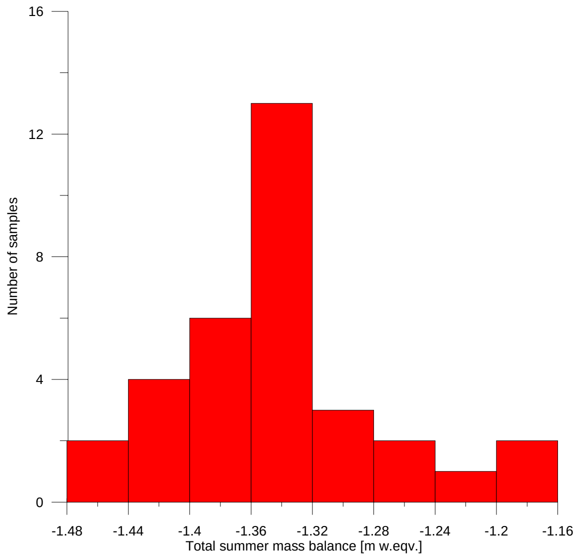

Figure 4.10 The graph displays the sensitivity of total summer mass balance to a random variable area weight factor.

To validate the result of the total summer balance I applied the ablation gradient approach using the same data. There I multiplied the specific summer balance in each elevation band by the hypsometry. The sum over the entire area of the glacier gives the total summer balance. Average specific summer balance where used if several stakes existed in a single elevation band. In ablation bands with no stake present, I used the neighbouring stake as representative. The results between the contour and the ablation gradient approach are -1.35 m water equivalent and -1.28 m water equivalent respectively with a difference of 0.07 m water equivalent representing 5 % of the total summer balance (fig. 4.11). The difference is a result of differential weighing of the area assign for the specific summer balance at that point between the two methods. Area means of specific summer balance are not representative for the specific ablation in each elevation interval due to the differential ablation pattern.

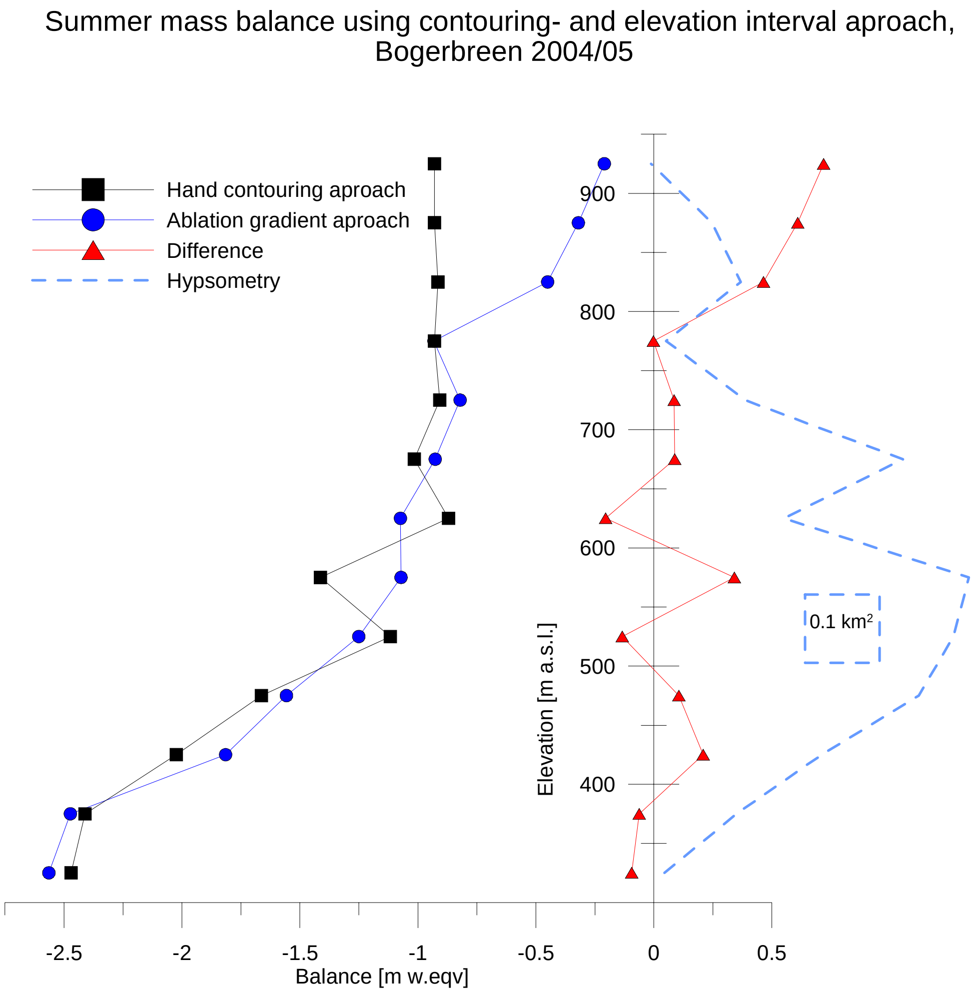

Figure 4.11 Plot showing the spatial difference of specific summer balances in each elevation interval for the contour approach and for the elevation interval approach.

Further enhanced is that discrepancy by extrapolation of the ablation gradient into areas. This is common practice for many mass balance related studies (Haefeli, 1962). On Bogerbreen the lowermost 300 – 350 m as well as the three upper bands 800 – 850 m, 850 – 900 m and 900 – 950 has no stake located within. I used a linear function following the trend of the 4 neighbouring stakes. Then the difference between the Contour- and the ablation gradient is even enhanced to -1.35 m water equivalent and -1.23 m water equivalent respectively representing 0.12 m water equivalent. That is 9 % of the total summer balance.

It is largely accepted that debris cover alters the ablation in a differential manner (Nicholson, 2004; Østrem, 1959). Mainly controlled by debris thickness it either accelerates or depresses the melt rate compared to that of clean ice. As results on Larsbreen, a nearby glacier show, if a thin veneer of debris is present, ablation can be enhanced by 20 % (Nicholson, 2004). If debris exceeds a certain threshold of a few cm ablation decreases compared to that of clean ice. However on Bogerbreen no detailed mapping of extend or thickness of debris has been undertaken and the effect is basically unknown. But observations indicate that debris is present and that the debris cover increases towards the margins of the glacier. Stakes are mostly located in central parts of the glacier and are therefore not very representative for in incorporating such effects of debris cover. It is therefore very likely that large areas of the glacier surface are affected by such processes uncertainties arise due to that fact. I calculated the possible effect by assuming that 5 %, 10 % and 15 % of the glacier is affected by debris and therefore enhance or decreases ablation by 20 % compared to that of clean ice. The results show that the calculated total summer balance is sensitive within ± 0.08 m water equivalent, ±0.16 m water equivalent, ±0.24 m water equivalent respectively. This constitutes to 18 %, 12 % and 6 % of the total summer balance.

As a summary of the analysis above, I argue that the total summer balance of -1.35 m water equivalent is certain between ± 0.24 m water equivalent or 18 % of the total summer balance. It is therefore most sensitive to variations in ablation due to effects of debris on the glaciers surface. Different hand contouring scenarios generate second largest uncertainties. Applying the ablation gradient approach on Bogerbreen to calculate total summer balance most likely underestimates the total summer balance. The difference between the two approaches is significant within the range of the uncertainties.

4.2.3 Net balance

In order to validate the result obtained for the total winter balance, I used the results from the elevation interval approach of the winter balance in conjunction with the ablation gradient approach of the summer balance. The resulting total net balance is -0.70 m w.eqv and therefore 0.10 m water equivalent less negative then the reference value of -0.80 m w.eqv calculated by using the preferred contour method (fig. 4.12). The difference constitutes to 13 % of the total net mass balance.

I estimate the maximum uncertainty for the net balance to be ± 0.24 m water equivalent corresponding to uncertainty of the summer balance. But due to the above mentioned validation I'm confident that the uncertainty of the net balance is somewhat lower approximately ± 0.20 m water equivalent or 25 % of the total net balance.

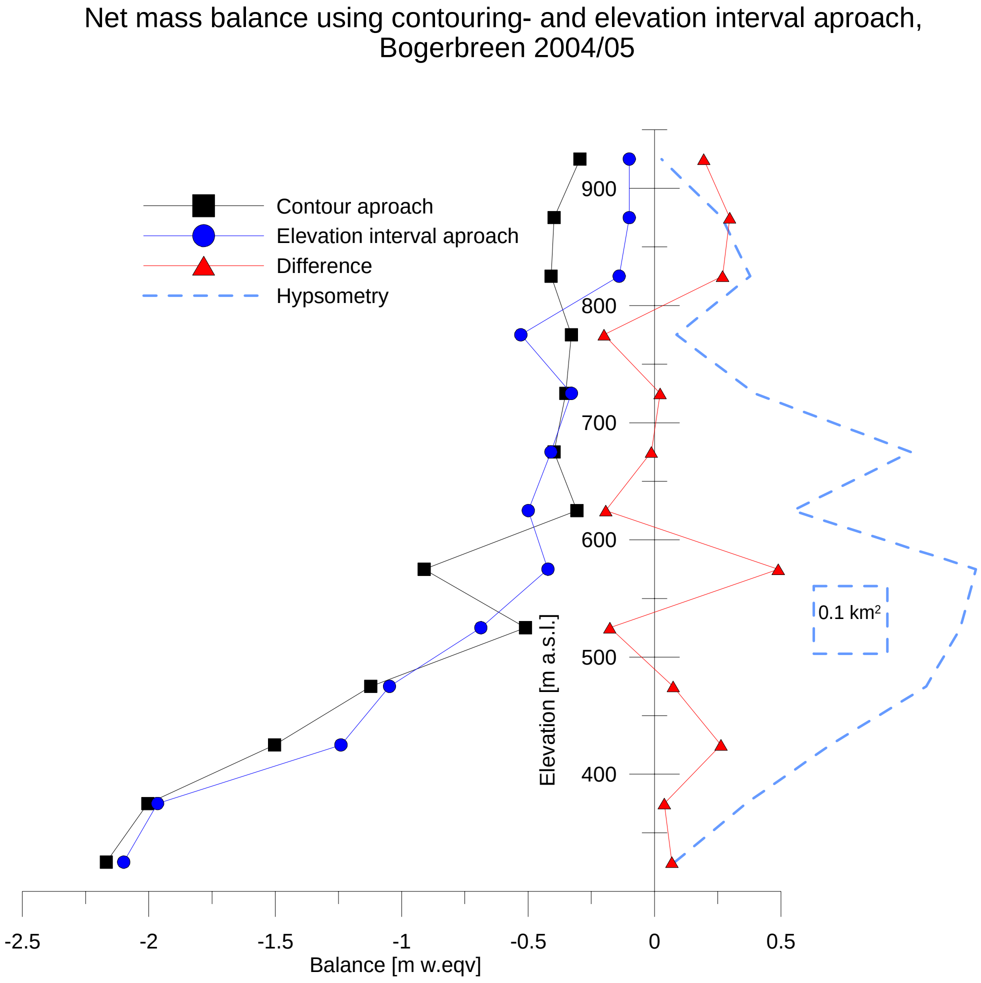

Figure 4.12 Plot showing the spatial difference of specific net balances in each elevation band for the contour approach and for the elevation interval approach.

4.3 Supplemental results

4.3.1 Temperature observations

Evaluating the temperature data collected on Bogerbreen lead to the following results. With help of the temperature record collected I determined mean temperature lapse rate (table 4.4), onset and termination of the ablation season 2004 and 2005 (table 4.5) and was able to extend the temperature record on Bogerbreen by correlation to the temperature record at Longyearbyen Airport (fig. 4.13). Mean summer air temperatures at the study site and Longyearbyen airport is presented in table 4.6. I find a good correlation between the meteorological station at Longyearbyen Airport and the correlation coefficient is $R^2 = 0.97$. Temperatures show a large variation during winter but little during summer season. Above cero temperatures events are frequent during winter season. In conjunction with the automatic obtained images the occurrence of summer accumulation can be pinpointed. The onset of the summer season 2005 is less pronounced compared to the onset of the winter season 2005.

Table 4.4 temperature lapse rate determined between temperature loggers at different altitudes on Bogerbreen and the meteorological station at Longyearbyen airport during one year period between 5th June 2005 and 4th of June 2006.

Sensor locations	Elevation interval [m a.s.l.]	Temperature laps rate [°C /100m]
Longyearbyen Airport to T1 temperature logger	28 - 335	0.98
Longyearbyen Airport to T2 temperature logger	28 - 478	0.96
Longyearbyen Airport to T3 temperature logger	28 - 707	0.95

Table 4.5 Onset and termination of the ablation season 2004 and 2005 for different altitudes derived from direct or extended air temperature record at Bogerbreen.

Description	Elevation 335 m a.s.l.	Elevation 478 m a.s.l.	Elevation 707 m a.s.l.
Termination of the summer season (balance year 2003/04), onset of the winter season (balance year 2004/05)	7th of September 2004	7th of September 2004	17th of September 2004
Termination of the winter season (balance year 2004/05), onset of the summer season (balance year 2004/05)	4th of June 2005	4th of June 2005	21st of June 2005
Termination of the summer season (balance year 2004/05), onset of the winter season (balance year 2005/06)	9th of September 2005	7th of September 2005	5th of September 2005

Table 4.6 Mean summer temperature (June – August) at different altitudes in 2005.

Sensor location	Elevation [m a.s.l.]	Mean summer temperature (June – August) [°C]
Longyearbyen Airport	28	5.1
T1 Bogerbreen	335	2.3
T2 Bogerbreen	478	1.2
T3 Bogerbreen	707	0.9

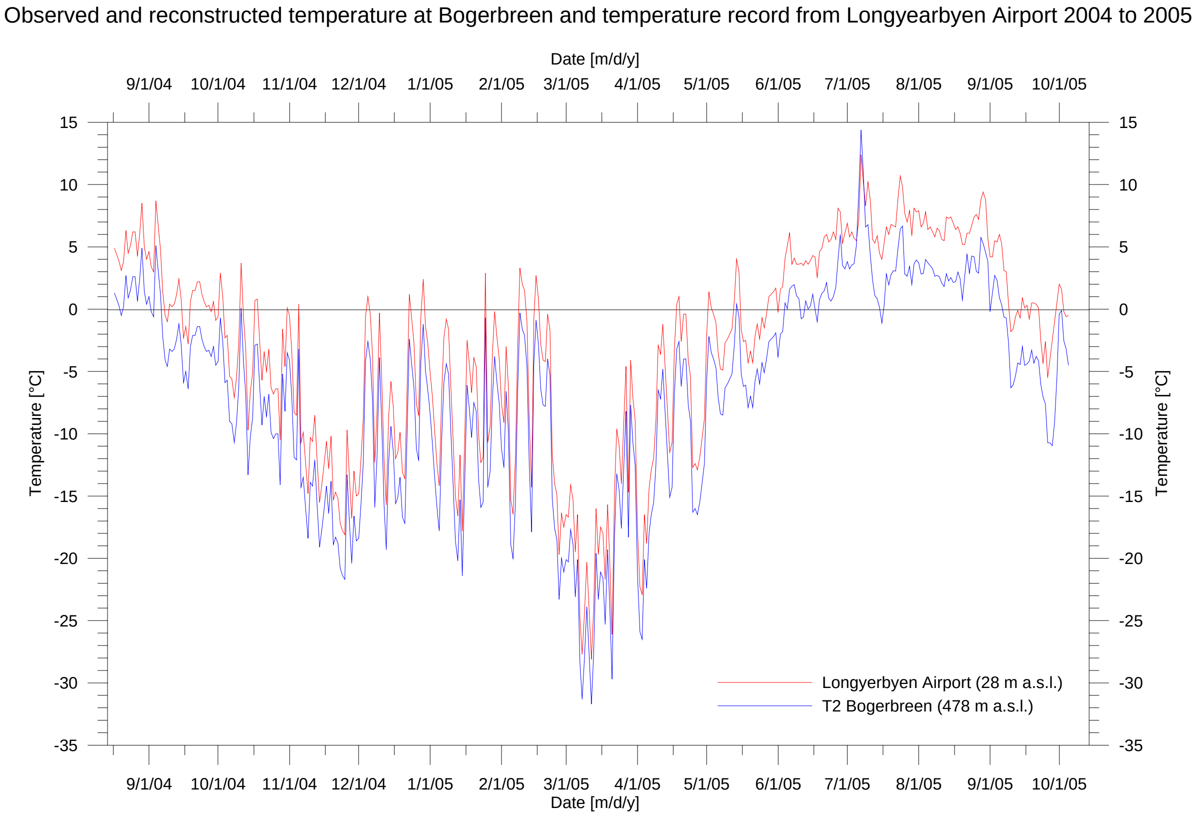

Figure 4.13 Temperature observations from Longyearbyen Airport plotted with temperatures at Bogerbreen. The Temperature on Bogerbreen where recorded from June 2005 until May 2006. Using temperatures from Longyearbyen Airport as a basis, Bogerbreens record could be reconstructed for the period August 2004 until May 2005. Air temperatures at Longyearbyen Airport from (eKlima, 2006)

4.3.2 Daily glacier monitoring

The snowline could be traced successfully using the images of the RDC during the balance year 2004/05. On 25 images the receding snowline could be identified and graphically mapped. However it was not possible to distinguish between bare ice and superimposed ice on the glaciers surface. The snow receded first from the north-eastern terminus on 16th of June and progressed along the eastern part of the glacier upward (fig. 4.14. a, b, c). The western side of the glacier remained snow covered until 18th of July. On 15th of July, summer snow precipitation covered the entire glacier. The snowline exceeded above the glacier limit on 7th of August 2005. On 31st of August only small patches of snow remained in the upper parts of the cirque (fig. 4.14. d). On 4th of September snow precipitation covered the upper cirque of the glacier while on 9th of September the entire glacier where covered with snow. This snow remained on the glacier for the following autumn and winter month. All remotely obtained digital images are given in the appendix, in a digital file format. Also included there is a “glacier-movie” with selected images of the remote digital camera.

Figure 4.14 The four images are taken from the remote digital camera located at Bogerbreen. The approximate position of the snowline is indicated by the red line. Images are taken during summer season 2005 on (**a**) 16th of June, (**b**) 30th of June, (**c**) 4th of July, and (**d**) 31st of August.

4.3.3 Base map generation

The generated high resolution DEM showed some inconsistent data in the area around the upper cirques. Unfortunately the problem could not be solved in a satisfying manner and therefore the low resolution DEM for Norwegian Polar Institute served as the foundation of the base map. The result is shown as topographic map (see fig. 2.4).

Bogerbreen hypsometry, representing the glacier state in summer 1990 is presented in figure 4.15.

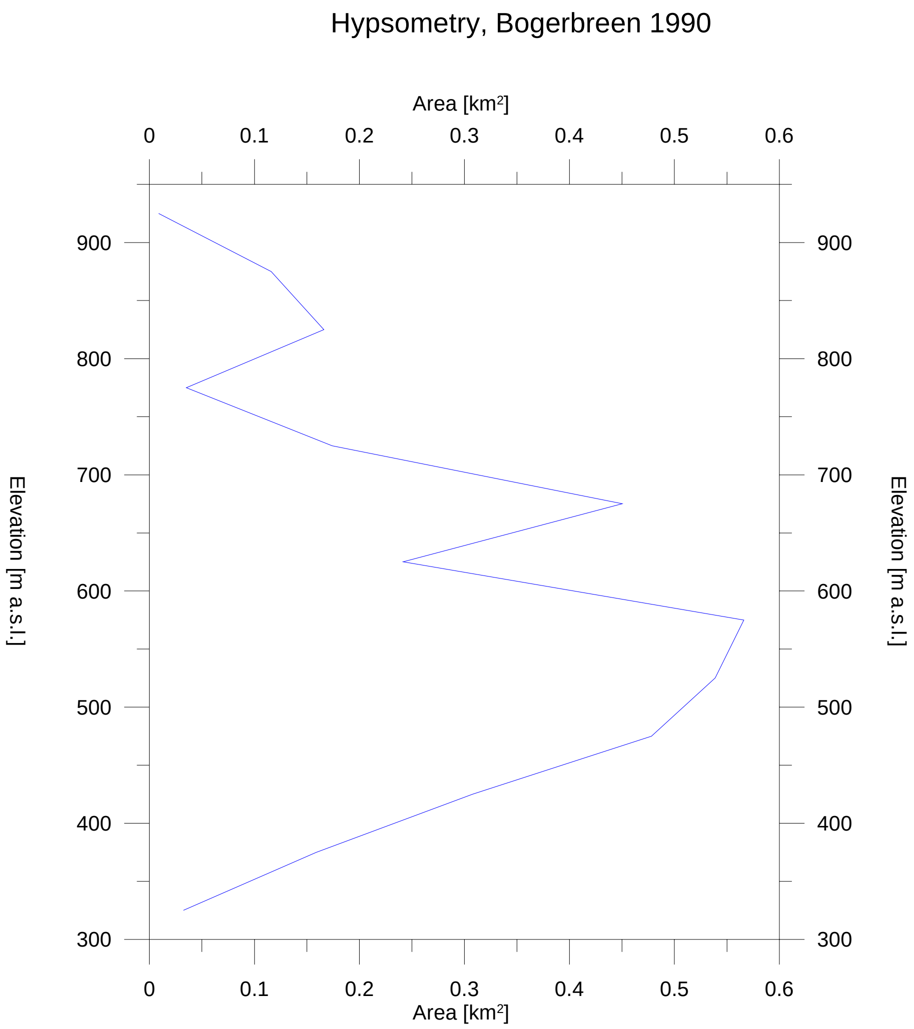

Figure 4.15 Hypsometry of Bogerbreens obtained from photogrammetrical survey of the 1990 summer surface.

On the other hand the consistent part of the high resolution DEM gave an independent data source to evaluate the low-resolution DEM. By comparing both DEM's, the RMSE for the *Z* coordinate was within 5.8m. For the glacier surface itself the RMSE decreased to 2.2 m. The RMSE between the DEM and the 10 ground control points measured on 1st of October 2005 were within 2.6 m. However, figure 4.16 indicates that the low-resolution DEM is tilted towards east or in fact that there is a horizontal shift towards the east since the error increases in sloping terrain. The generated high resolution DEM is attached in the appendix.

Knowing that the glacier is a highly variable system the base map should be up to date data for hypsometrical calculations on the glacier (Kaser *et al.*, 2003; Østrem and Brugman, 1991). However down wasting instead of receding dominates many Svalbard glaciers (Humlum *et al.*, 2005) therefore its likely that the total surface area of Bogerbreen has changed relatively little.

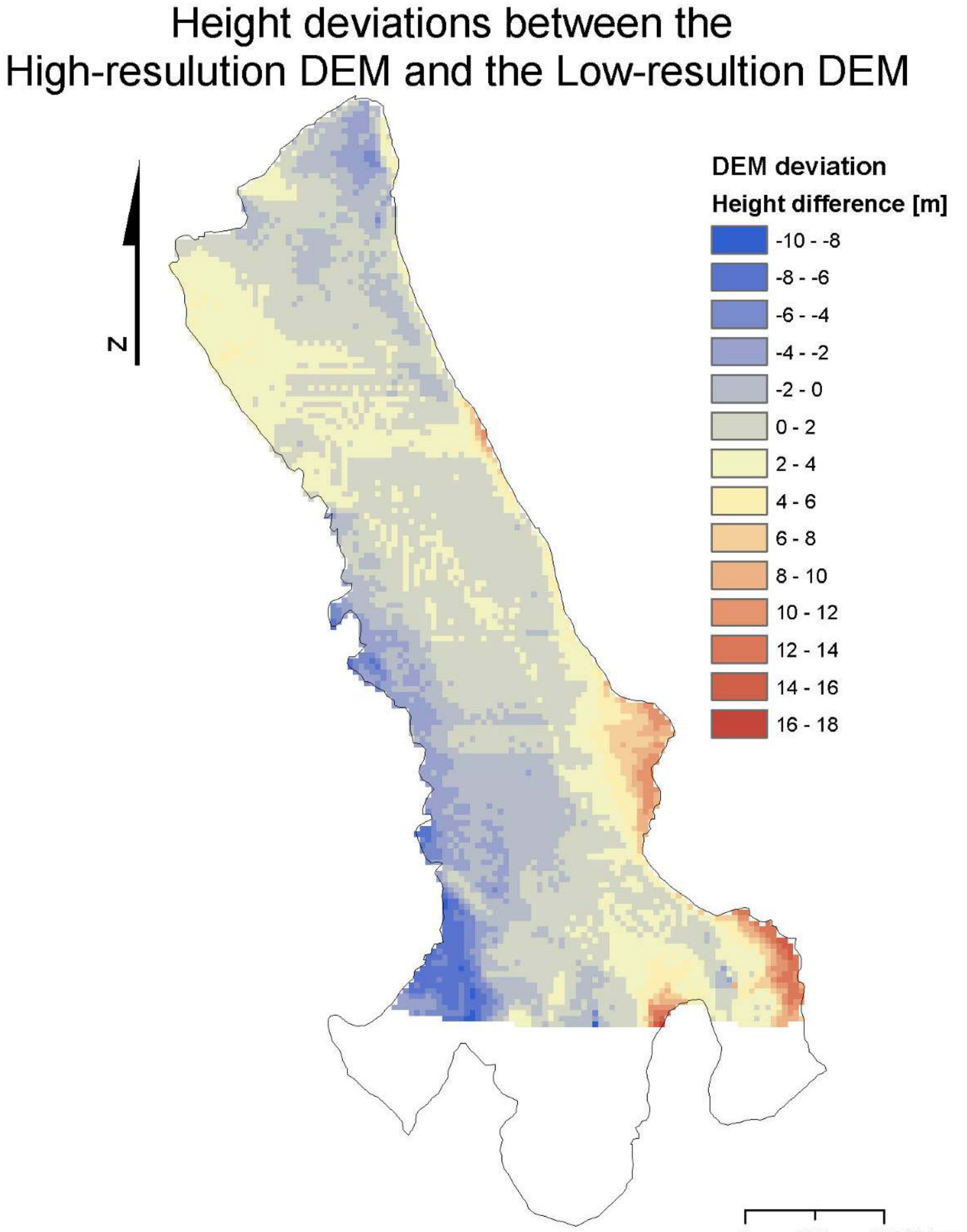

Figure 4.16 Map showing the height deviation from the low-resolution DEM derived from 1:50.000 scale images compared to the height-resolution DEM derived from 1:15.000 scale images. Both aerial surveys where flown within two weeks in 1990.

The result from the orthorectification, were six individual orthoimages with a pixel resolution of 1 m covering the glacier catchment (fig. 4.15). The absolute error has been evaluated by identifying ground control points surveyed In September 2005. A total of 6 points where identified and the error proofed to be within the pixel resolution of 1 m in all the orthoimages except one. Due to an error within the rectification process, orthoimages number 4 has a shift of 3 m towards the east. The six individual orthoimages are attached in the appendix.

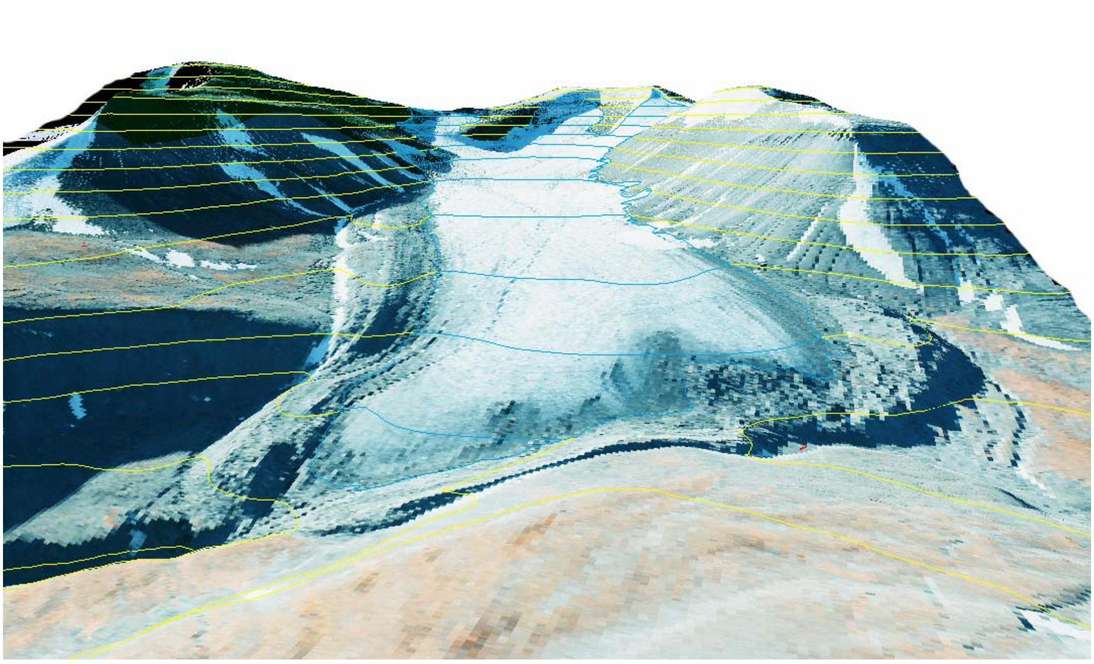

Figure 4.17 A mosaic of 6 generated orthoimage draped over the low resolution DEM from Norwegian Polar Institute. The image is directed south. The displayed contours are spaced with 50 m intervals.

4.3.4 Ice surface velocities

Average, annual ice velocity during the measured period is 0.81 m a^{-1} (fig 4.18). Measured velocities range from 1.38 m a^{-1} at the highest stake down to 0.49 m a^{-1} at the lowest stake. The velocities decrease almost linear down glacier. The standard error is estimated to be within 1 decimetre. The coordinates for the stake locations are given in the appendix, in a digital file format.

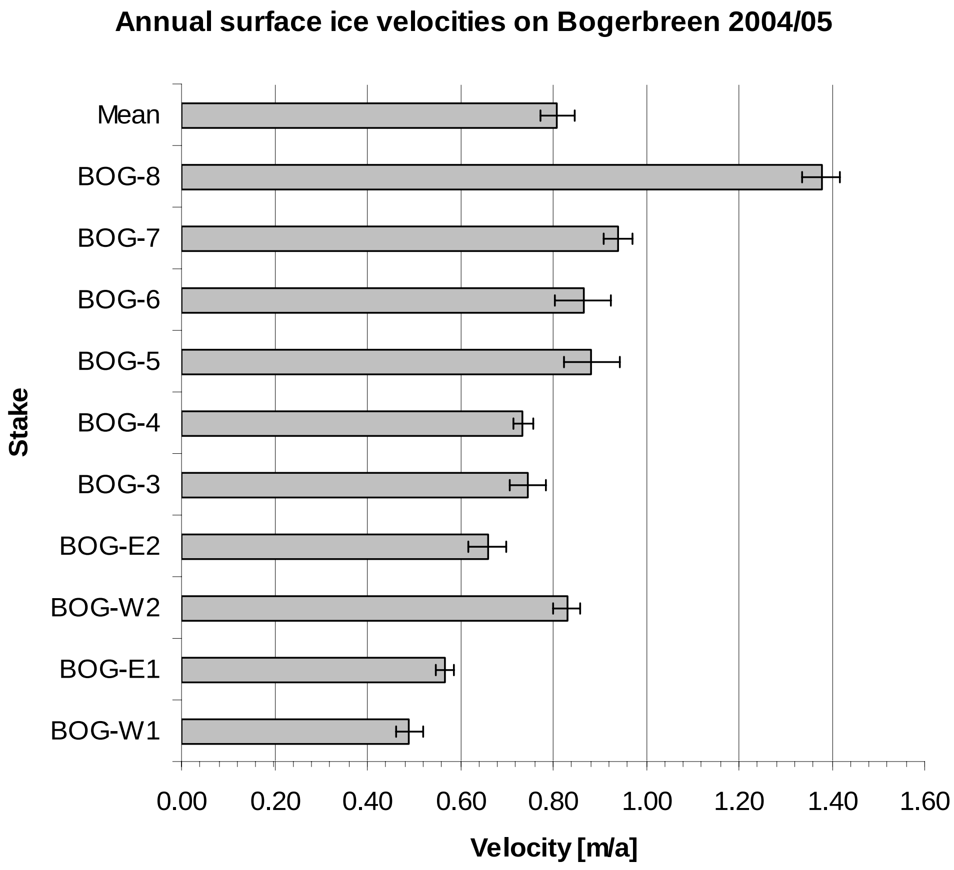

Figure 4.18 Annual surface ice velocities on Bogerbreen during 2004/05 derived from DGPS measurements. The error-bars represent one standard deviation.

4.3.5 Mean geodetic balance 1990 to 2004

The mean annual net mass Balance on Bogerbreen between the balance years 1990/91 and 1003/04 were clearly negative, calculated to be -0.65 m water equivalent . The mean surface elevation change over the entire glacier was 9.30 m. Figure 4.19 clearly indicates that the entire glacier has thinned. Largest elevation changes were observed at the terminus decreasing towards the head of the glacier. It was also found that in the middle part of the glacier in elevations between 400 m a.s.l. and 600 m a.s.l., the glacier has thinned more along the eastern margin (fig. 4.20). Raw date such as the kinematic DGPS survey and the high resolution DEM is attached in the appendix.

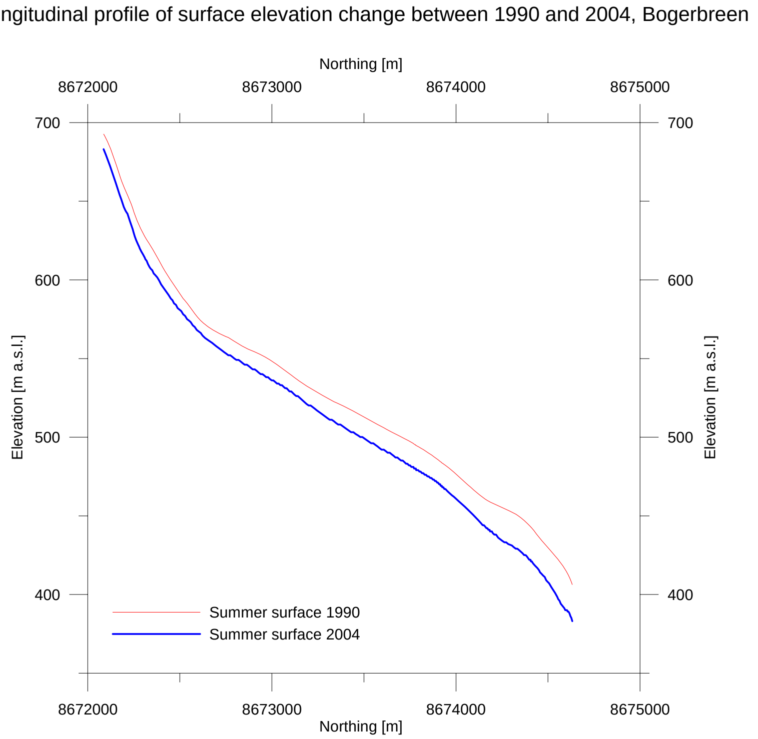

Figure 4.19 The figure shows the elevation change of the glacier surface along a longitudinal profile.

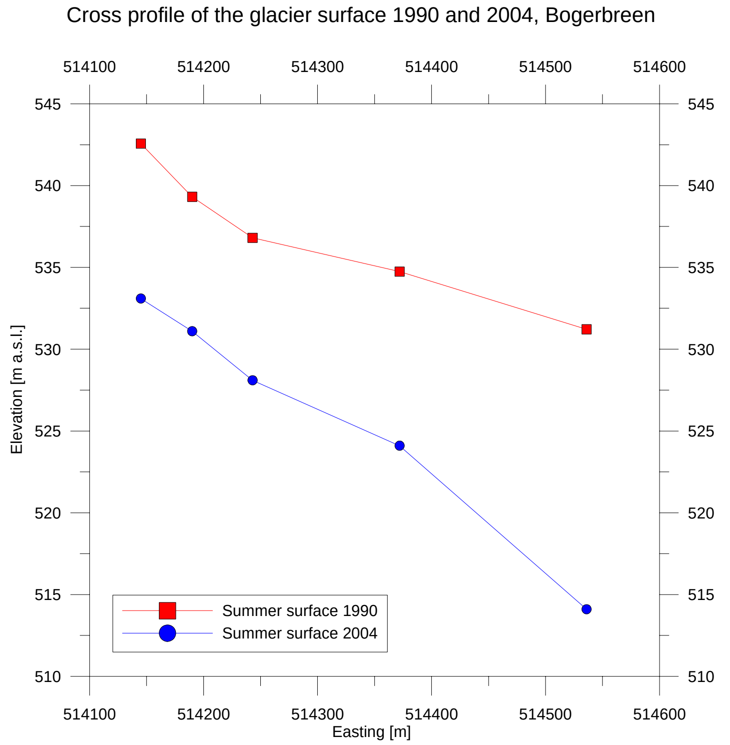

Figure 4.20 The figure shows the elevation change of the glacier surface along a cross profile. The profile is located approximately at stake BOG-5. Note that

the change on the eastern margin of the glacier is almost twice the change of the western margin.

Unfortunately the value yields a large degree of uncertainty. The largest error of ± 0.22 m water equivalent, constituting 32 % of the net balance, is introduced by the uncertainty of the low resolution 1990 DEM and the positions from the kinematic DGPS profiling. A somewhat smaller uncertainty is introduced by extrapolation into higher and lower elevation intervals. By applying a wide range of possible ablation rates in those areas the mean geodetic balance is sensible within 0.06 m water equivalent or 9 % of the calculated balance. It is most likely that after aerial survey has been conducted on 29th of July 1990 ablation has proceeded. This indicates that the above mentioned balance is most likely underestimated to some degree.

An attempt was made to validate the above mentioned result using independent survey data. Therefore I used the high resolution DEM from 1990 and stake positions obtained with differential GPS in spring 2005. I subtracted the stake height above the ice surface for each of the stakes and extracted the elevation from the high resolution DEM for each of the stakes X, Y coordinates. A hand-adjusted ablation curve were fitted and extrapolated into lower and higher elevations. The so obtained balance was calculated to be -0.68 m water equivalent. Even this approach yields similar uncertainties; it is very close to the first estimate and therefore strengthens the validity of the result.

5. Discussion

5.1 Mass balance on Bogerbreen

5.1.1 Winter balance

In the following chapter I will first discuss spatial distribution of specific winter balance on Bogerbreen followed by the controls on the winter mass balance.

The spatial distribution of the winter balance is largely heterogeneous. Bogerbreen has no or only a little pronounced specific winter balance gradient as a function of elevation, but there is s a pronounced lateral gradient. Snow distribution is fairly homogeneous as a function of elevation (fig. 5.1). It is very likely that the spatial distribution of winter balance is largely controlled by redistribution of snow at the surface. Redistribution of snow on Svalbard is considerable. Solid precipitation, high relative wind speeds during winter, and lacking higher vegetation are the major causes for that (Jaedicke, 2002; Humlum, 2002).

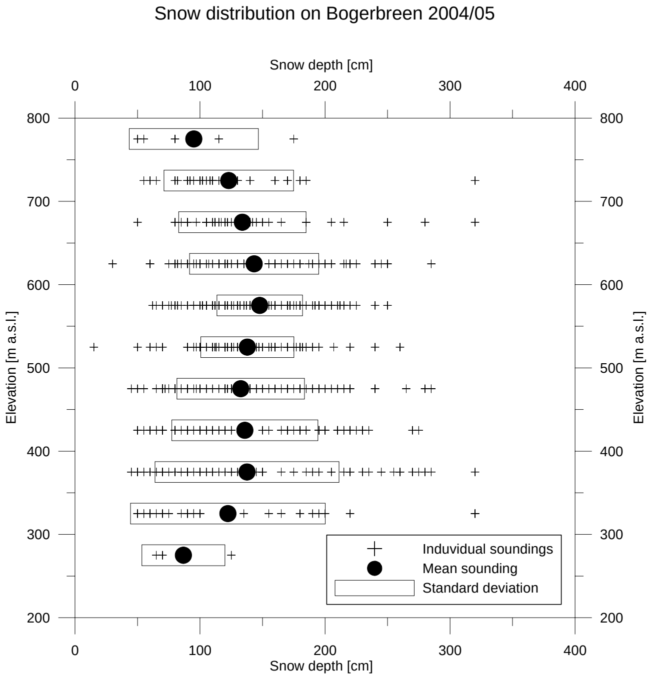

Figure 5.1 Snow depths as a function of elevation on Bogerbreen.

Bogerbreen is elongated in the dominant regional winter wind direction for wind speeds exceeding 5 m s $^{-1}$ (fig. 5.2. b). Air is mainly draining from the upper part of the glacier towards the terminus. From there it is directed north eastwards, into Endalen and Adventdalen due to lesser air resistance (see also map fig. 2.4). This is according to observations I made. While in upper part of Endalen snow erosion (Sastrugies) suggests high wind speed, erosion in upper Fardalen seldom takes place, suggesting lower wind speeds. Due to redistribution processes, the snow will form a pattern, which is linked to a combination of topography and wind direction. Wind speeds on Bogerbreen seem to be highest along the eastern central flowline of the glacier leading to effective erosion of snow there. Along the western margin where topographic friction suggests slower wind speeds deposition takes place.

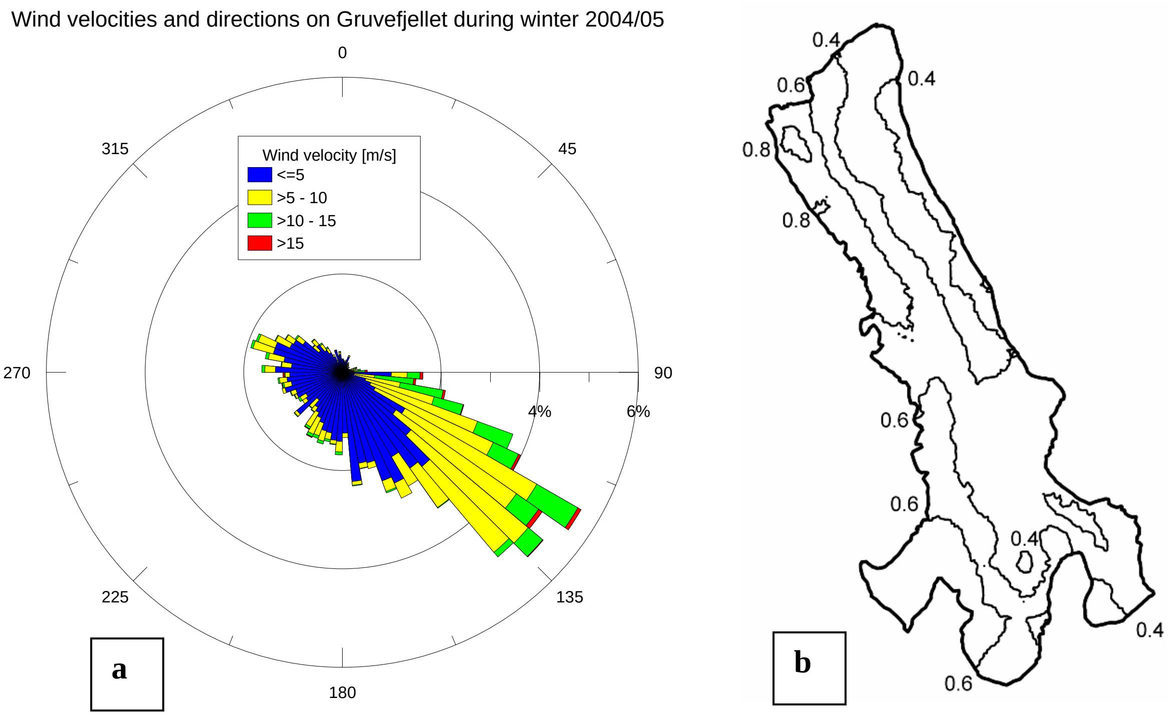

Figure 5.2 The regional winter wind velocities and direction is recorded at Gruvefonna approximately 4 km north ob Bogerbreen (**a**).The map (**b**) displays the spatial distribution of specific winter mass balance on Bogerbreen.

For many smaller valley glaciers such as Bogerbreen the amount of accumulation increase with elevation due to topographic controls (Orvin, 1991). On Bogerbreen the effect seems to be altered to a large degree by redistribution of snow from the upper part to the lower part. The distribution of specific winter mass balance in 1982/83 (Guskov and Troitskiy, 1985) and 1983/84 (Guskov and Troitskiy, 1987) were very similar to the one observed 2004/05 (fig. 5.3). This suggests that the distinct pattern observed in 2004/05 seems to be rather persistent with time.

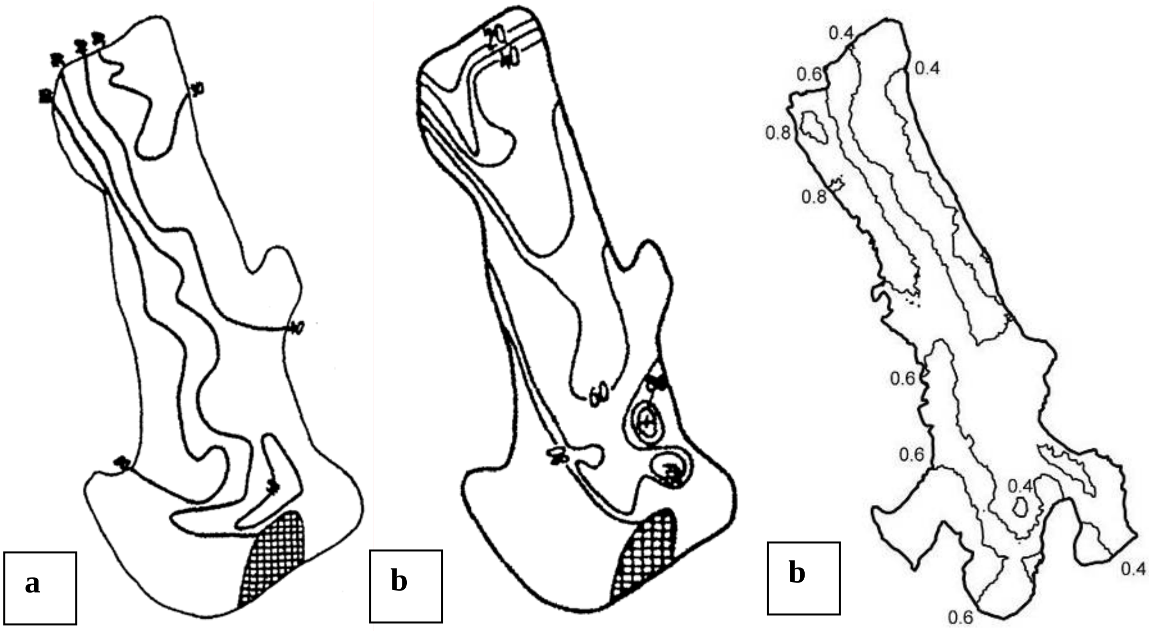

Figure 5.3 Spatial distribution of winter mass balance on Bogerbreen. North is up. Winter balance 1982/83 (**b**) in g cm^{-2} (Guskov and Troitskiy, 1985). Winter balance 1983/84 (**b**) in g cm^{-2} (Guskov and Troitskiy, 1987). Winter mass balance 2004/05 (**c**) in m water equivalent.

Traditional methods, such as snow probing measurements used in this study, incorporate several mass balance components into one measurement and make it difficult or impossible to determine the individual quantities for each of the components. To find the main control on winter mass balance on Bogerbreen I will start with the less contributing.

Avalanche is a common source of accumulation on many mountain glaciers (Benn and Evans, 1998; Patterson, 1994), but conditions on Svalbard are mostly uninvestigated. In the balance year 2004/05 the accumulation due to avalanches proved to be insignificant. Nevertheless most surrounding slopes and headwalls on Bogerbreen have critical inclinations which favour snow avalanches. One indication that avalanches did occur on Bogerbreen is presented in figure.5.4. One avalanche deposit is visible on the aerial picture S 90 5467, taken on 16th of August 1990. The conditions in 2004/05 might not represent the general condition on Bogerbreen. If meteorological conditions favour avalanches, this can lead to an increase of the accumulation area and avalanches might contribute to the winter mass balance in a significant manner.

Figure 5.4. Aerial photograph showing an avalanche deposits along the western margin on Bogerbreen indicated by the circle. North is up. Photo: Norwegian Polar Institute, scale: 1:15000, 16.August 1990, S1990 5467.

The formation of autumn superimposed ice changes only very little on the overall picture of the specific- or total mass balance during the balance year 2004/05. Superimposed ice formed mostly on the little inclined ice surface between stake BOG-7 and BOG-5 and around stake BOG-W2. Certain estimates of superimposed ice are very difficult to obtain due to its large spatial variation (Hagen Jon Ove and Reeh, 2004; Konig *et al.*, 2002). Especially the measurement made on stake BOG-W2 might not represent the occurrence of superimposed ice very well. Especially in this area closer to the terminus the glacier surface is more inclined and with a hummocky appearance making it very difficult to correctly estimate the formation of superimposed ice. Even though there little influence of superimposed ice formed in the balance year 2004/05 that should not lead to the assumption that it might have insignificant influence in all years. Hagen *et al.* (1991) states that superimposed ice can contribute up to 30 % to the net balance on high arctic glaciers. The fact that large areas of Bogerbreen were covered with superimposed ice in autumn 2006 underlines the statement further.

Snow contributed most to the winter balance and incorporates precipitation, surface rime and wind transported snow. Surface rime is mostly very low compared to quantities of precipitation and redistributed snow (Patterson, 1994). Even though it is very difficult to quantify the contribution from those factors there are indications that redistributed snow plays a significant roll on Bogerbreens winter mass balance.

Redistribution of snow on Svalbard is considerable as already mentioned above. In western and central parts of Svalbard, winter precipitation is insufficient to maintain the present-day mass balance of local glaciers, according to Jaedicke and Gauer (2005). During redistribution events, snow is eroded and transported from plateaus on Nordenskiöld Land and deposited in lee positions and on glaciers (Chinn *et al.*, 2005). Conclusions are verified by results from a snow drift model. In Nordenskiöld Land, accumulation areas of glaciers coincide with areas of low wind speeds and therefore areas of snow deposition (Jaedicke and Gauer, 2005). Bogerbreen is located in such a leeside position relative to the dominant winter wind direction (see also map fig. 2.4). It is therefore very likely that wind drifted snow add significantly to the winter mass balance beside precipitation.

For the Bogerbreen record, correlation between winter mass balance and winter precipitation at Longyearbyen Airport is low with correlation coefficient of $R^2 = 0.61$. I identify two major causes for that 1) Redistribution is significantly altering the winter mass balance on Bogerbreen and, 2) Precipitation observations are unreliable. The latter fact is due little catch efficiency resulting in serious measurement problems for precipitation measurements on Svalbard (Førland and Hanssen-Bauer, 2000; Hansen-Bauer *et al.*, 1996). As can bee seen in fig. 5.5 correlation between winter mass balance and winter precipitation at Longyearbyen Airport is somewhat better in the period 1975/76 to 1984/85, than for the balance year 2004/05. This might be an indication of large interannual fluctuations of redistributed snow. On the other hand, the measuring problems at the metrological station might be the cause for the variability in correlation.

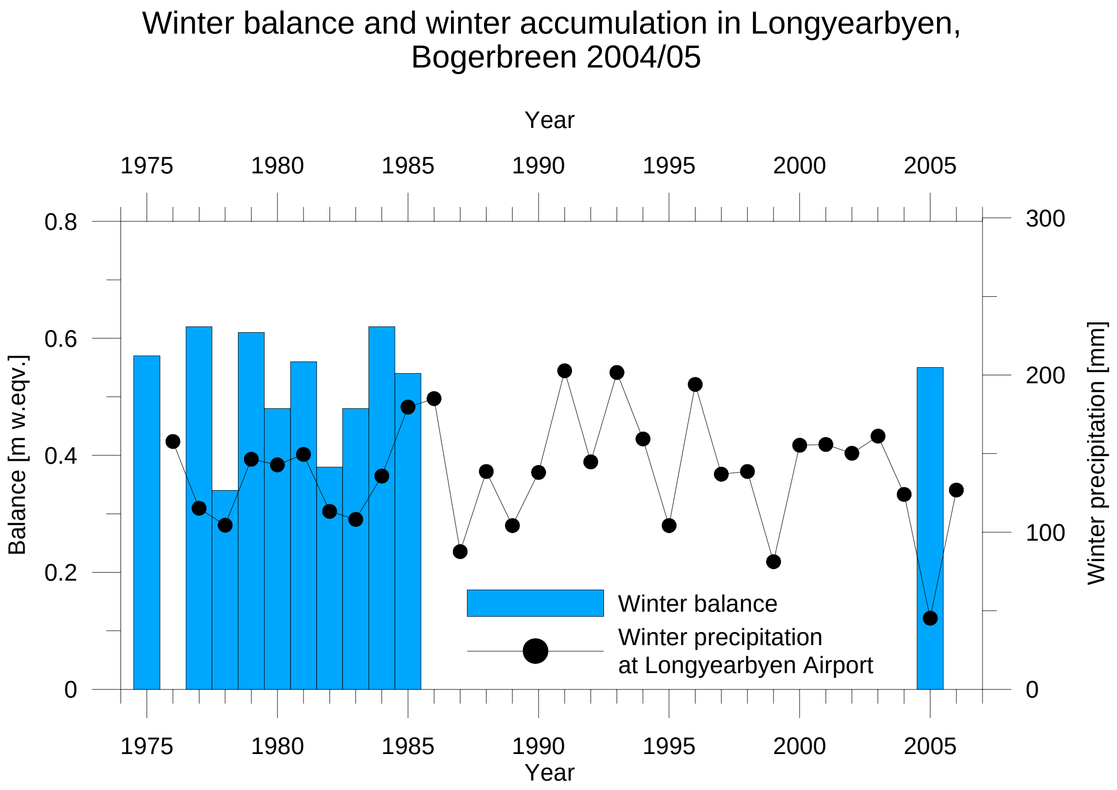

Figure 5.5 The plot visualizes the total winter accumulation for the available record on Bogerbreen and the winter precipitation observed at Longyearbyen Airport. The winter mass balance values between 1975 and 1985 are findings from (Kotlyakov, 1985).

Summarizing the above mentioned arguments, topography and prevailing wind direction- and speed seem to be the principal controls on the spatial distribution on specific winter balance on Bogerbreen. It is likely the wind transported snow plays a significant roll contributing to Bogerbreen winter mass balance beside precipitation. Avalanching and internal accumulation contributed only little to the overall winter mass balance in 2004/05.

5.1.2 Summer balance

In this chapter I will first focus on the spatial distribution of the summer mass balance and its controls in the balance year 2004/05, followed by a discussion using the existing mass balance record between 1974/75 to 1984/85.

Glacier mass loss processes are melting followed by run off, evaporation but also internal ablation and basal melt. However, traditional surface mass balance measurements does not account for the latter two (Patterson, 1994). Low flow rates, ice thickness below 100 m and the lack of icings may suggest that basal melt can be neglected. Except for the mass loss of snow by wind, ablation is the result of energy supplied to the ice or snow surface. Energy is derived either by radiation or directly from the air due to convection and condensation. Near the surface, temperatures decrease with barometric pressure, due to that thermal energy in a mass of air is proportional to its density (Benn and Evans, 1998). There is little doubt that heat transfer from the air is a function of air temperatures (Patterson, 1994; Schytt, 1967). Bogerbreen showed the local laps rate is inversely correlated to the summer balance with a correlation coefficient of $R^2 = 0.87$ (fig. 5.6). Such a linear relationship has been also proposed by Schytt (1967). Therfore I explain the observed gradient of decreasing summer mass balance with increasing elevation mainly to that fact.

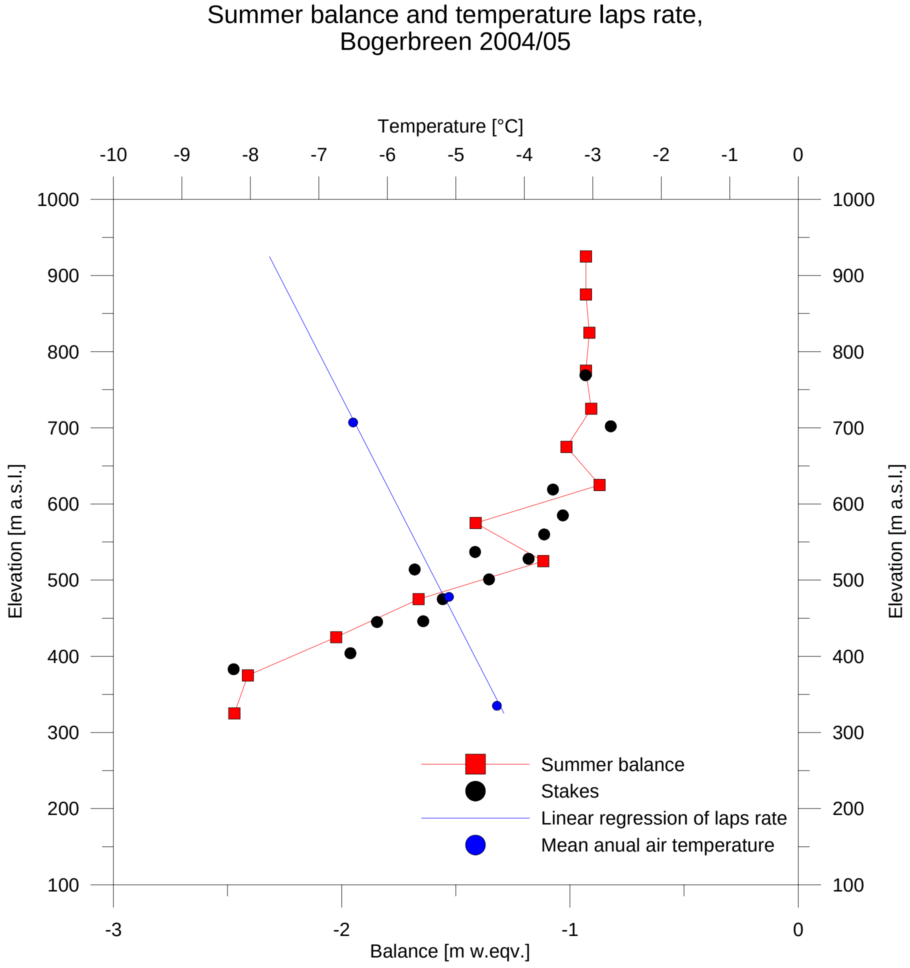

Figure 5.6 The graph displays the inverse relationship between local temperature laps rate and summer balance on Bogerbreen

However not all the spatial variability observed on Bogerbreen can be explained due to heat transfer from the air. In cold regions the main control on ablation is short-wave radiation which is only weakly dependant on elevation (Kuhn, 1984). Modification of the linear gradient can be caused by a variety of processes. Those can be altitudinal variations in cloudiness and humidity, proximity to rock walls, the amount of shading, glacier aspect and variations in albedo (Oerlemans and Hoogendoorn, 1989; Kuhn, 1980). Without being able to support my presumptions with measurements, I argue that the most likely cause for the lateral gradient are differences in albedo and amounts of direct radiation. Due to the small glacier size possible causes such as variations in cloudiness and humidity are assumable small. But the topography surrounding Bogerbreen is highly variable, so that the proximity to rock walls, which influence the amount of available long-wave radiation and sensible heat transfer, could also contribute.

Differential ablation due to debris cover is the common most cause of non-linear ablation gradients on many glaciers (Nakawo and Rana, 1999; Benn and Evans, 1998; Østrem, 1959). Nicholson (2004) studied the influence of debris cover on the nearby glacier Larsbreen. While thin debris cover up to just a few cm enhance melt up to 20 % compared to clean ice, further increase in debris thickness lead to suppressing melt rates considerably until melt is almost entirely stopped. Debris-cover has been observed on Bogerbreen and thicknesses vary significantly between thin veneers to several decimetres. Its very likely that especially at the glacier margins, where debris is abundant, melt is enhanced. That is supported by results from the stake readings across the glacier. However the process does not explain the cause of the strong lateral west-east summer balance gradient existing 2004/05.

Also differences in surface albedo caused by spatial distribution of snow and ice during melt season can cause non-linear ablation gradients (Patterson, 1994; Schytt, 1967). As images of the remote digital camera reviled, early in the melt season the eastern margin becomes free of snow while along the western margin snow duration is much longer (see also fig. 4.14). As soon as snow melts and bare ice gets exposed surface albedo is reduced (Patterson, 1994). Albedo is reduced along the eastern margin since snow vanishes there first during ablation season. This is leading to increased melt rates on the eastern side compared to the conditions along the western margin. These effects can also contribute to the longitudinal gradient as mentioned earlier in that chapter. The time transcressive character of the snowline during spring and autumn, as well as snow accumulation during summer will most likely enhance the longitudinal gradient of ablation on Bogerbreen. Summer accumulation occurred on 15th of July covering the entire glacier. While lower parts where snow free already on 18th of July, upper parts remained snow covered until 31st of July as automatically obtained images reviled.

Another likely cause for the distinct lateral gradient is the shading effects hindering the effectiveness of direct radiation. Bogerbreen is surrounded by mountains towards east, south and west. It is very likely that direct radiation on the glaciers surface is increasing from west to east explaining the distinct lateral gradient. But it would also lead to increasing direct radiation towards the north of the glacier. In other words direct radiation is increasing towards lower elevations on Bogerbreen due to its aspect. Therefore it might add to the above mentioned longitudinal gradient caused by convection and albedo, increasing the gradient further.

The use of the cumulative positive degree days was found to be a useful parameter linking meteorological observations and summer balance observation on glaciers on Svalbard (Szafraniec, 2002; Hagen and Liestøl, 1990). For the entire balance record including measurements from Guskov (1988), Gokhmann *et al.* (1988), Guskov and Troitskiy (1987), Guskov and Troitskiy (1985), I find a reasonably good correlation between total summer balance and positive degree days both in Longyearbyen and for the reconstructed temperature record on Bogerbreen (fig. 5.7). The correlation coefficient increases even more if we sum positive degree days between May and September instead of traditionally used June to August positive degree days. The correlation coefficient for a sample of 11 is $R^2 = 0.78$. This is in good agreement with findings of Hagen and Liestøl (1990) for similar sized glaciers in the north east of Spitsbergen.

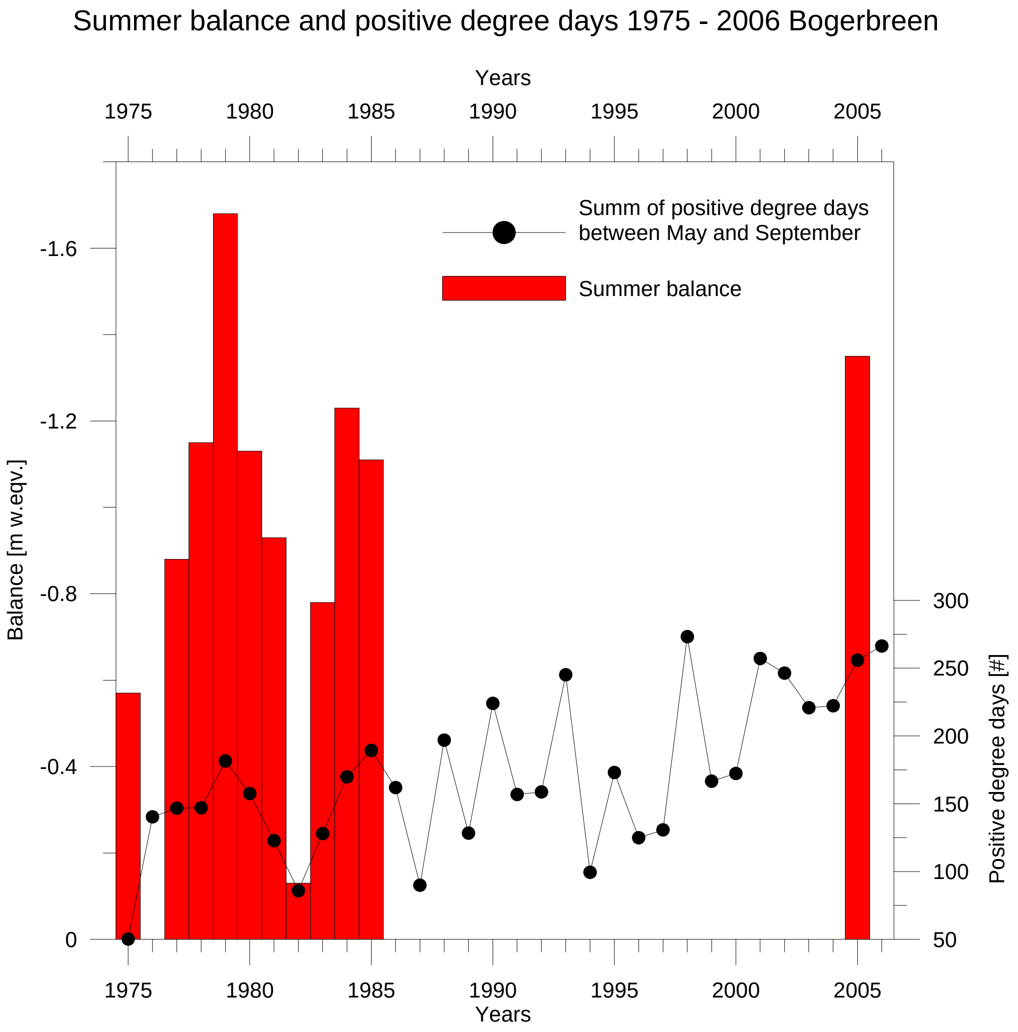

Figure 5.7 The plot visualizes the accumulative sum of positive degree days at Longyearbyen Airport between 1975 and 2006 and the total summer balances for the available record on Bogerbreen. The total summer balance values between 1975 and 1985 are findings from (Kotlyakov, 1985).

I can summarize the above discussed issues as following. It is likely that heat convection by air is the cause for the distinct summer balance gradient as a function of elevation. Both shading effects on direct radiation and effects caused by different albedo on the surface (snow, ice and debris) add to the gradient. Lateral gradients can be explained by differences in shading and surface albedo. Topographic shading and the spatial distribution of snow and consequently variations in surface albedo are important controls leading to the asymmetrical summer mass balance pattern on Bogerbreen. Evidence both from surface surveys and aerial pictures suggests that this distinct asymmetrical pattern has been existing on Bogerbreen during the last few decades. However glacier dynamics can not be excluded as a likely cause for the asymmetrical pattern. The cumulative sum of positive degree days at Longyearbyen Airport can be used as a reasonable approximation of summer balance on Bogerbreen.

5.1.3 Net balance

The spatial pattern of net mass balance as observed in the balance year 2004/05 shows consistency to earlier published maps showing the spatial distribution of net mass balance on Bogerbreen (Guskov, 1988; Gokhmann *et al.*, 1988; Guskov and Troitskiy, 1985) (fig. 5.8). It’s striking how consistent the asymmetry of the spatial distribution for net mass balance is for the years 1977/78, 1982/83 and 2004/05. Assuming that albedo effects is the major control on the net balance asymmetry, the similar spatial distribution of net mass balance in 1977/78, 1982/83 and 2004/05 might indicate that spatial pattern of snow at the start of the ablation season has been fairly stable.

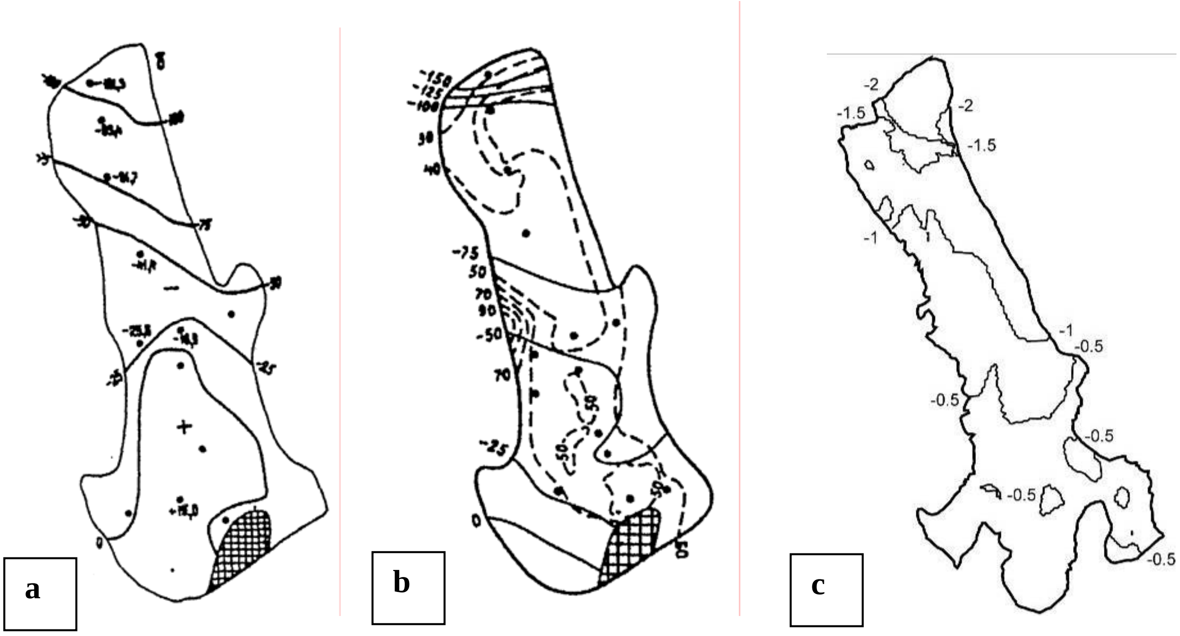

Figure 5.8 Spatial distribution of net mass balance on Bogerbreen. North is up. Net balance 1977/78 (**a**) in g cm^{-2} (Guskov, 1988). Net balance 1984/85 (**b**) in g cm^{-2} (Gokhmann *et al.*, 1988). Net mass balance 2004/05 (**c**) in m water equivalent.

Results, from the surface surveys, strengthen the above mentioned lateral balance asymmetry. The surface lowering at the eastern margin in the middle part of the glacier has been twice as much as on the western margin for the period 1990/91 – 2003/04 (see also fig. 4.20). I can not clearly state to what extend glacier dynamic plays a significant role here. Less accumulation along the eastern tributary could also be the cause of the enhanced rate of surface lowering along the east. On the other hand, snow distribution in 1982/83, 1983/84 and 1984/85 is very similar to 2004/05 (see fig 5.3). Therefore the difference in albedo is likely the cause of the enhanced ablation along the easy compared to the west. Evidence of enhanced melt along the eastern margin, is also found by studding aerial pictures from the Norwegian Polar Institute. Photos taken in summer 1960 (S60 7245) and 1990 (S90 5467) show both bare ice exposed on the terminus and along the eastern margin at the middle part of Bogerbreen. Also the change of supraglacial channel routing between 1960 and 1990, as observed from the above mentioned aerial pictures, indicates that the eastern margin of the glacier has lowered relative to the western side (fig. 5.9). This might indicate that the distribution of snow at the start of the ablation season were similar to the one observed 2004/05 over the last few decades.

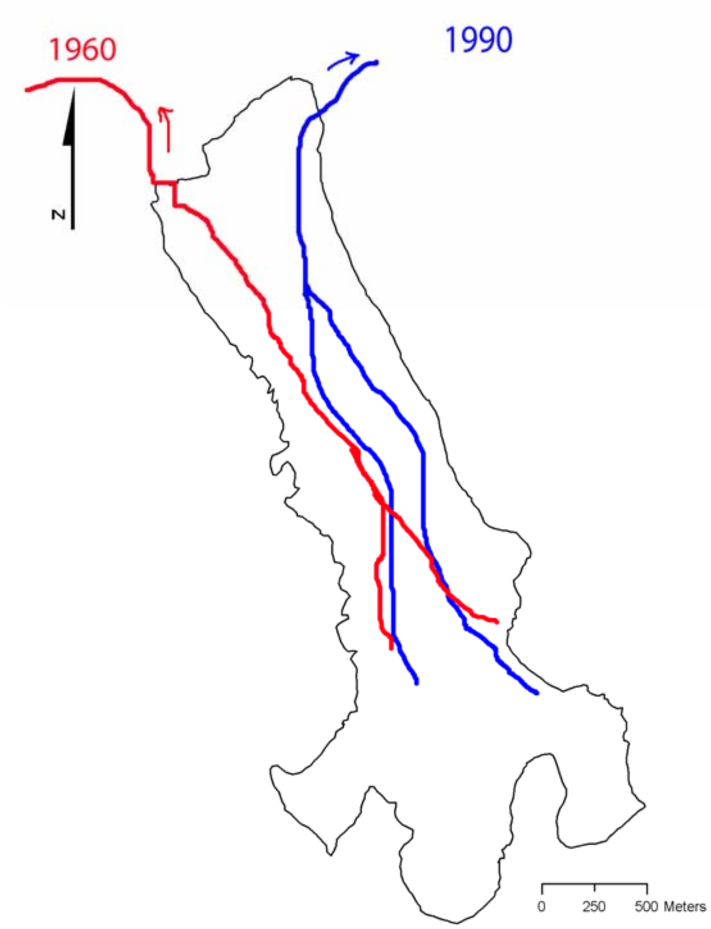

Figure 5.9 Change in supraglacial channel routing observed from aerial pictures 1960 (S60 7245) and 1990 (S90 5467) on Bogerbreen.

Bogerbreen net mass balance is strongly dominated by the summer mass balance. The correlation coefficient between summer mass balance and net mass balance on Bogerbreen is R^2 = 0.97. This is consistent with the widely accepted opinion that mass balance of continental glaciers are more influenced by summer mass balance (Janson, 1999; Fountain and Vecchia, 1999; Kuhn, 1980). While winter mass balance show only minor variations, summer balance has a larger inter annual range. The same is valid for the resulting net mass balance (fig. 5.10). It is therefore clear that Bogerbreen net mass balance is very sensitive to changes in the ablation regime during the summer season. However, findings of UNESCO (1970) revealed how strongly linked summer runoff is controlled by the spatial distribution of winter accumulation. Observations on Bogerbreen 2004/05 support their studies. The distinct pattern of winter accumulation at the start of the summer season seems to effect the distribution of the summer balance distribution and consequently the net mass balance. Nevertheless winter accumulation processes are only little understood yet (Munro, 2000).

Indicated by the negative net balance, Bogerbreen was not in balance with the present climate in the balance year 2004/05. Bare ice where exposed from the terminus up to the headwalls without showing signs of superimposed ice accumulation. Also mean geodetic net mass balances between 1990/91 to 2003/04 and the existing mass balance record from Kotlyakov (1985) between 1975/76 to 1984/85, having mainly negative mass balances, give further indication that Bogerbreen hasn’t been in balance since the last 3 decades (fig. 5.10).

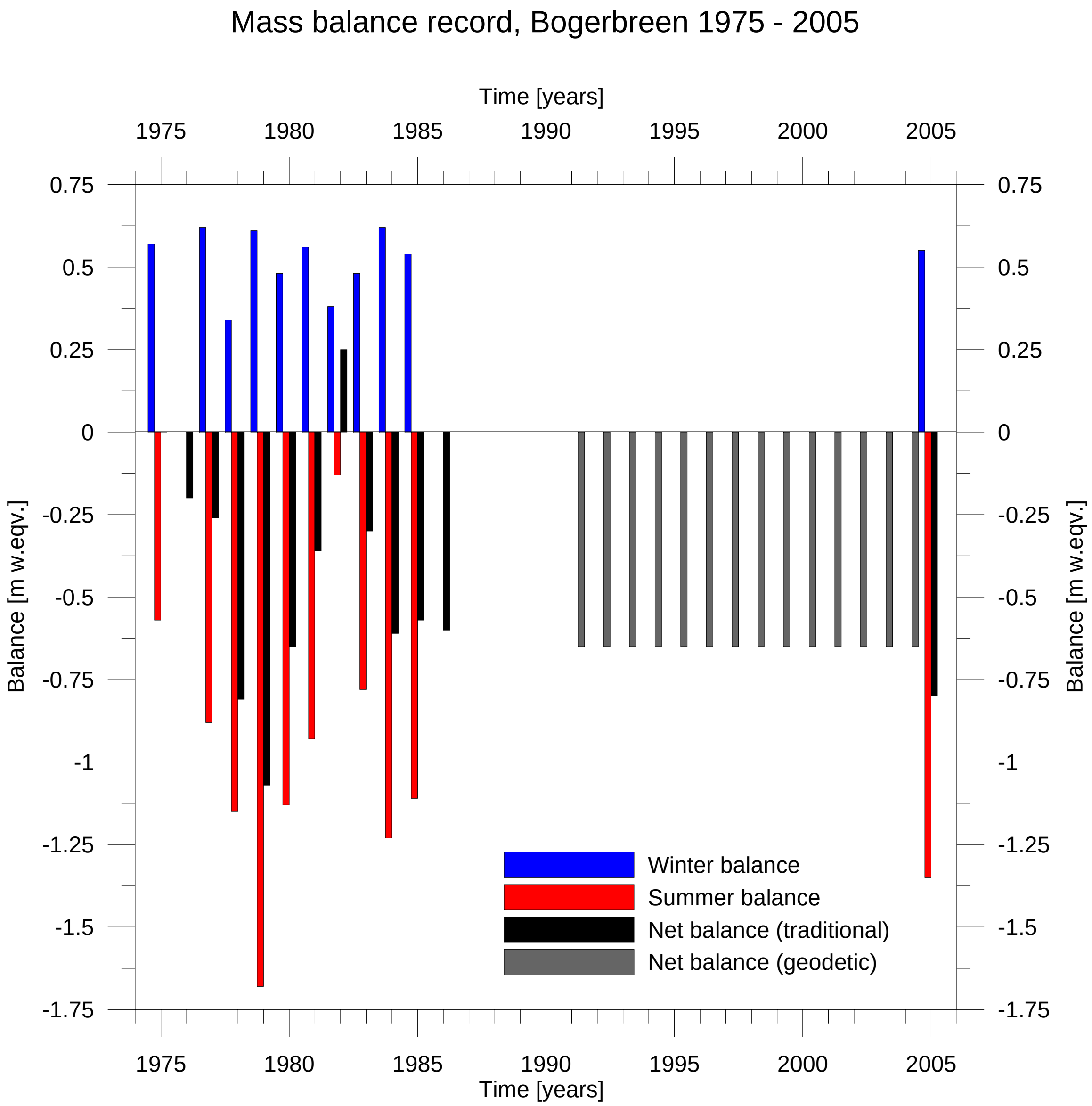

Figure 5.10 Graph showing the record of mass balance measurements performed on Bogerbreen. Data between 1974/75 to 1984/85 are obtained from REF 33,34,35. Note, the period between 1990/91 to 2003/04 is only represented by mean net mass balances obtained using the geodetic method.

The ELA is an important parameter that links the climatic conditions of ablation and accumulation with glacier behaviour (Benn and Evans, 1998; Patterson, 1994; Furbish and Andrews, 1984). As mentioned above, in 2004/05 the ELA has exceeded the upper glacier limit and even those of the surrounding peaks. Extrapolation of the net mass balance gradient would indicate an ELA at approximately 1000m (fig. 5.11). Estimated ELA for zero net mass balance on Bogerbreen is 540 m a.s.l. (Hagen, 1993), (fig. 5.12). This fact strongly supports the argument that Bogerbreen is not in balance to the present day's climate.

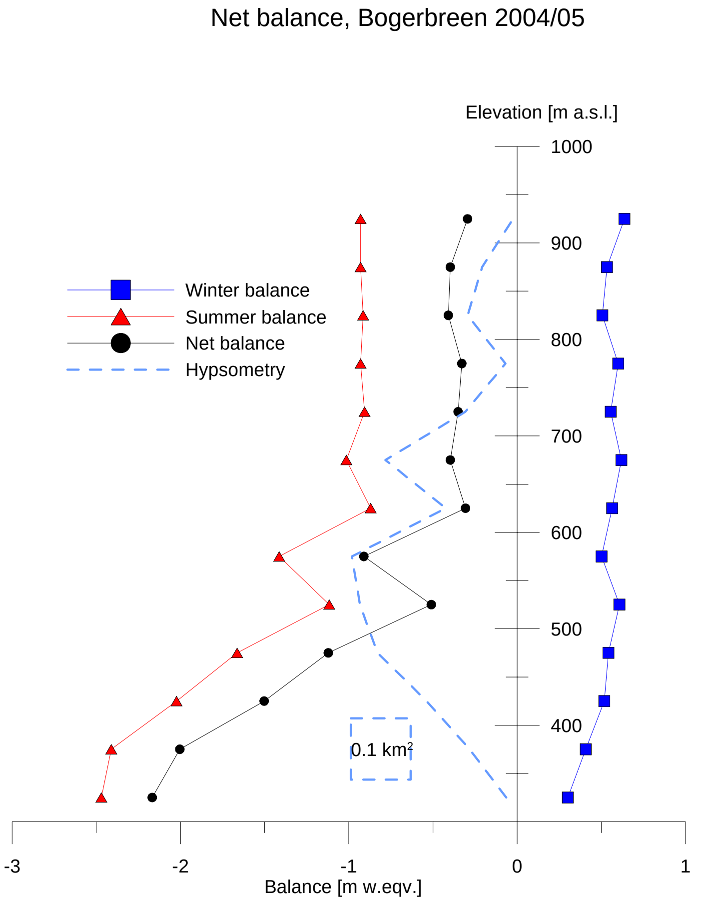

Figure 5.11 Winter- , summer- and net mass balance on Bogerbreen 20004/05 versus elevation.

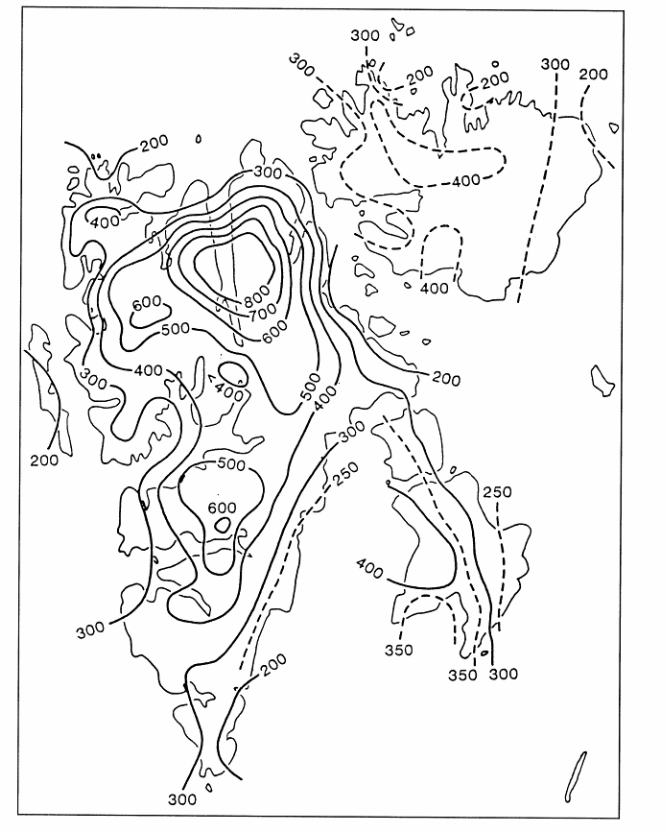

Figure 5.12 Estimated ELA on Svalbard for the condition of zero net mass balance. From (Hagen, 1993).

5.1.4 Error analysis

Several issues concerning mass balance methods and there uncertainties have been discussed in the workshop on methods of mass balance measurements and modelling, Tarfala, Sweden August 10-12, 1998. Uncertainties of traditional methods widely remain unknown (Østrem and Haakensen, 1999; Fountain *et al.*, 1999). Within this study I tried to determine the level of uncertainty of my measurements mostly by calculating their possible response to the balance obtained, or in other words calculate the sensitivity of the mass balance value to possible uncertainty range of measured parameters.

I argue that the total winter balance has the lowest level of uncertainty (± 0.08 m water equivalent or 15 % of the total winter mass balance). For that I see basically two reasons: 1.) The widespread large sampling density and 2.) The favourable conditions on Bogerbreen 2004/05 (little snow depth, thin ice lenses). Janson (1999) concluded on

Storglacären that if only total winter mass balance is of interest probing network can be reduced if a marginal error can be tolerated. However, detailed measurements are necessary in order to give a representative image of the spatial distribution of winter mass balance, due to its large variation. Largest uncertainties arise from snow density measurements. Ice lenses are frequent especially at the lower half of the glacier and can cause erroneous probing to a false summer surface.

For the summer mass balance uncertainties exceed those of the winter balance (± 0.24 m water equivalent or 18 % of the total summer mass balance). Most likely the less dense sampling of the stake measurements compared to winter accumulation measurements is the major factor contributing to the summer balance uncertainty. Fountain and Vecchia (1999) concluded that for valley glaciers, where mass balance variations are dominated by elevation, 5 to 10 stakes along the centre line are sufficient if covering the elevation range of the glacier. As discussed above, there are indications that on Bogerbreen the relation between mass balances as a function of elevation is altered due to variations in albedo, direct radiation, and spatial pattern of winter accumulation just to name some. The occurrence of summer accumulation, in the form of snow, leads to underestimation of melt. The images from the remote digital camera proved very valuable in reporting such events and their extent.

Calculations showed that traditional contouring proved to be more representative method compared to the ablation gradient approach. Due to the lateral gradient the common ablation gradient approach lead to underestimation of summer mass balance on Bogerbreen of - 0.13 m water equivalent or 10 % of the total summer mass balance. Even though the discrepancy was significant only within the limits of uncertainties, Bogerbreens summer mass balance is sensitive towards the method applied to calculate the balance. Therefore I argue that a rather dense network is needed initially due to the large spatial variations. As soon the pattern has been taken into account the network might be reduced. Similar findings were made on South Cascade Glacier and Meclure Glacier by Fountain and Vecchia (1999). Especially higher elevations on Bogerbreen were not represented by stakes mostly due to practical reasons (proximity to steep headwalls and avalanching). But, I found that especially in those unsempeld regions errors can occur due to hand- or computer aided extrapolations. This is in accordance with studies made on Storglacären (Janson, 1999).

Maximum uncertainties of the net mass balance where within ± 0.20 m water equivalent or 25 % of the total net mass balance. Base maps used for area calculation should be up to date (Janson, 1999; Østrem and Brugman, 1991). Here on Bogerbreen the effect of the changing topography or in this case hypsometry has not been emphasized much. However, down wasting instead of receding dominates many Svalbard glaciers therefore its likely that area distribution has changed only little (Humlum *et al.*, 2005; Etzelmueller *et al.*, 1996).

The mean net mass balances obtained using the geodetic method yielded larger uncertainties compared to the traditional method (± 0.22 m water equivalent or 34 % of the total mean net mass balance). The poor quality of the low resolution DEM seems to be contributing most to the uncertainty. Nevertheless, the geodetic method proved useful for determining mean net balances especially over longer periods such as decades (Østrem and Haakensen, 1999).

In general the presented uncertainties are somewhat less than the studies by REF 8 show. I argue that the large sampling density of both winter and summer balances is the major cause for that discrepancy. Nevertheless the sums of 42 field days spent on Bogerbreen made me feel well acquainted with the glacier and therefore also helped reducing uncertainties.

I want to make clear that all the correlations given within this analysis might not be statistically significant. Typically, the level of significance is controlled by both the sample size (number of measurements) and the nature of the distribution (Davis, 2002). Unfortunately in many cases the number of measurements is low and the distribution often unknown.

Past mass balance records were mainly published in Russian language. While Hagen and Liestøl (1990) finds good agreement with former soviet measurements it is not clear if discrepancies exist due to the use of different methodology. Translations of the papers could not clearly reveal the methods used to determine winter- and summer mass balance. Literature indicates that summer mass balances have bean measured using the hydrological method between summer 1975 and 1985, using gauge measurements. Against that stays filed evidence. I found cables, weights and tetrahedrons exposed on the surface of Bogerbreen on 12th of august 2005 (fig. 5.13). This kind of installation is used as an alternative to stakes, to measure ice ablation (Østrem and Brugman, 1991). This method is very similar to the stake method. The hydrological method has a very different approach consequently leading to different values. Nevertheless, most likely measurements are in accordance with present day measurements even though detailed information concerning methods on Bogerbreen could not be reviled.

Figure 5.13 Tetrahedron, cable, weight and the marker of the installation to measure glacier surface ablation. Such installations were found on several locations during a field visit on 12th of August 2005.

5.2 Mass balance on Svalbard

There are at present only six glaciers on Svalbard where traditional mass balance investigations are performed on an annual basis: Austre Brøggerbreen, Midre Lovénbreen, Linnébreen, Kongsvegen, Hansbreen and Bogerbreen (fig. 5.14), (table 5.1). Data for the glaciers monitored by the Norwegian Polar Institute were easily accessible (Dr. Jack Kohler, pers. comm.) while data for Hansbreen could unfortunately not be obtained in time to be included in the present study.

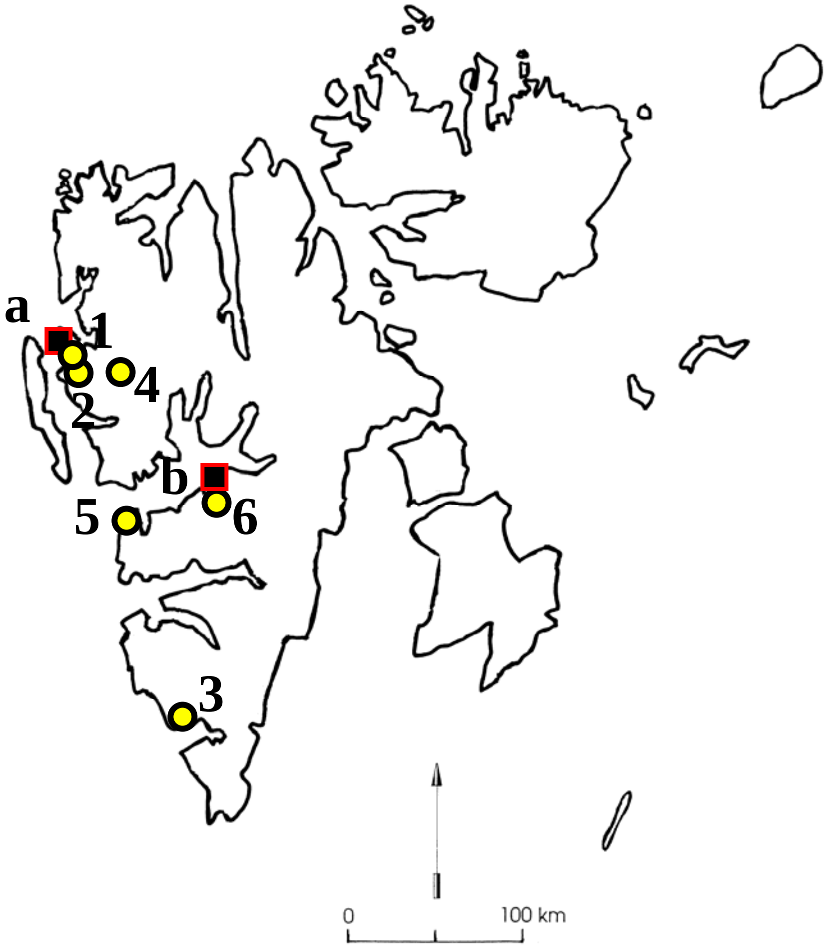

Figure 5.14 Map of Svalbard. The dots represent location of glaciers where annual mass balance measurements are performed at present. Glacier names as follows: (**1**) Midre Lovénbreen, (**2**) Austre Brøggerbreen, (**3**) Hansbreen, (**4**) Kongsvegen, (**5**) Linnébreen and (**6**) Bogerbreen. The squares locate metrological stations: (**a**) Ny Ålesund, (**b**) Longyearbyen Airport. Modified from (Hagen, 1993).

Table 5.1 Some selected glacier key values of the compared glaciers. Data for glaciers marked with (*) from (Hagen, 1993).

d [#]	Name	Length [m]	Area [km^2]	Elevation range [m a.s.l.]	Aspect
1	Midre Lovénbreen*	4.8	5.59	40 - 650	NE
2	Austre Brøggerbreen*	6.0	11.80	40 - 600	N
3	Hansbreen*	15.6	64.00	0 - 550	S
4	Kongsvegen*	27.0	189.00	0 - 1050	NW
5	Linnébreen*	3.85		150 – 550	E
6	Bogerbreen	4.1	3.30	325 – 925	N

5.2.1 Winter mass balance

Comparison of total winter balances 2004/05 show that mentioned glaciers have very similar winter mass balance (fig 5.15). The individual winter balances are mainly within the uncertainties of the measured balances. Linnébreen has highest winter mass balance mainly due to the proximity to the precipitation source. Precipitation is smallest for the area where Bogerbreen is located, compared to other glacier monitoring sites (fig. 5.16). At the same time Bogerbreen is located at higher elevation. It is likely that increased precipitation as a function of elevation and deposition of redistributed snow is adding to Bogerbreen's winter mass balance, making it similar to the other mentioned glaciers. Wind blown snow has been reported as a significant contributor on winter mass balance for glaciers in central Svalbard (Humlum *et al.*, 2005; Hodgkins *et al.*, 2005; Humlum, 2002; Jaedicke *et al.*, 2000).

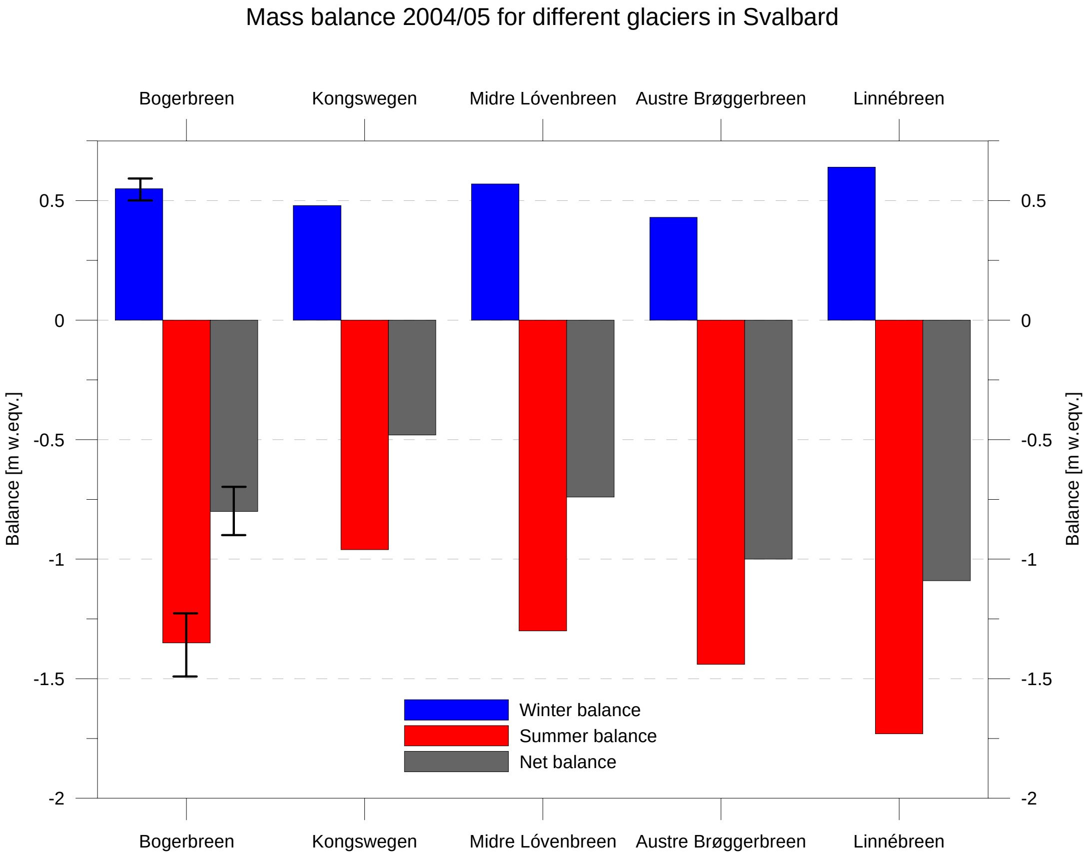

Figure 5.15 Total winter- summer and net mass balances for the different monitored glaciers on Svalbard. Error bars for Bogerbreen represent uncertainties for the measurement.

Figure 5.16 Precipitation in Svalbard [mm a^{-1}]. The map is mainly based on mass balance measurements of selected glaciers on Svalbard. From (Hagen, 1993).

Considering the entire mass balance records, Bogerbreen has a 25 % smaller winter balance compared to Linnèbreen, Austre Brøggerbreen and Midre Lovénbreen. Then the influence of continentallity is more pronounced than it is for the balance year 2004/05. The correlation coefficients for winter mass balance and winter precipitation on Bogerbreen is $R^2 = 0.61$. A small correlation comparable to the one for Bogerbreen was found on Midre Lovénbreen and Austre Brøggerbreen with $R^2 = 0.63$ (Hagen and Liestøl, 1990). There Hagen and Liestøl (1990) concludes that the unreliable precipitation measurements are the cause for the small correlation. On Bogerbreen I argue that both unreliable precipitation measurements and the effect of snow deposited by wind on the glacier contribute to a small correlation.

5.2.2 Summer balance

Total summer balance for the selected glaciers show a larger range than total winter balances (fig. 5.15). For Bogerbreen, Midre Lovénbreen and Austre Brøggerbreen total summer balances are similar within the levels of uncertainties. They also share similar glacier hypsometry and size but are located at different geographical setting and elevations (see fig. 5.14 and table 5.1). The continental setting of Bogerbreen suggests larger annual temperature amplitude and consequently higher summer temperatures. Mean summer temperatures (Mai – September) at Longyearbyen Airport and Ny Ålesund differed only little in 2005, 4.7 °C and 4.2 °C respectively. It is very likely that the temperature difference is compensated by the lapse rate, since Bogerbreen is situated at higher elevation. Under the assumption that turbulent heat fluxes play a major role, this might cause similar magnitude summer mass balances.

Even though Linnébreen is similar size to Bogerbreen, Midre Lovénbreen and Austre Brøggerbreen (table 5.1) it shows the most negative summer mass balance. The proximity of Linnébreen to relatively warm water masses passing along the west of Spitsbergen and their influence of turbulent heat fluxes might be one possible explanation. At the same time it can not be excluded that local variations in cloud cover, albedo, aspect or shading affect ablation compared to the other glaciers.

The comparison between Bogerbreen and Kongsvegen shows a different result. Kongsvegen has not only less negative total summer mass balance but it also has a very different hypsometry and glacier size compared to Bogerbreen. Kongsvegen is 25 times the size of Bogerbreen and has the bulk of its area around 700 m a.s.l. compared to 500 m a.s.l. at Bogerbreen. Reduced ablation in higher elevations, over larger glacier areas, led to reduced total summer balance of Kongsvegen.

5.2.3 Net balance

Total net mass balance for the six monitored glaciers vary similar manner than summer mass balances (fig. 5.15) All compared glaciers had negative net mass balances indicating that none of the glaciers were in balance state 2004/05. Austre Brøggerbreen and

Linnébreen show larger negative net mass balances indicating higher turn over rates compared to Bogerbreen. Midre Lovénbreen, but especially Kongsvegen, had less negative net mass balances compared to Bogerbreen. This is mainly a consequence of reduced summer mass balance. I see the main cause in the nature of the hypsometry difference between the two glaciers as discussed above.

The strong control of summer mass balance on net mass balance at present on Bogerbreen has also been observed on other glaciers on Svalbard. Considering the so far existing mass balance records, correlation coefficient for summer and net mass balance on Bogerbreen, Austre Brøggerbreen, Midre Lovénbreen and Kongsvegen are $R^2 = 0.97$, $R^2 = 0.88$, $R^2 = 0.83$ and $R^2 = 0.89$ respectively. The strong relationship between summer and net mass balance on Bogerbreen is most likely due to the somewhat more continental setting of Bogerbreen compared to the other mentioned glaciers.

5.3 Sensitivity, trend and future response of Bogerbreen

Future climate warming has been predicted to be more pronounced in the Arctic than in lower latitudes (Dowdeswell and Hagen Jon Ove, 2004). Benn and Evans (1998) outlines the importance of area/altitude distribution of glaciers with respect to glaciers mass balance sensitivity to climate changes. On Svalbard the present ELA is very close to the elevation of the bulk of the glacier area (Hagen *et al.*, 2003b). Due to the character of the hypsometric distribution, little change of the ELA is needed to cause a large effect on the net mass balance. Therefore Hagen *et al.* (2003a) concludes that the glaciers on Svalbard are rather sensitive to climate change. Bogerbreen shows a very similar picture (fig. 5.17). Actual ELA is up to 100m higher than those estimated for zero mass balance conditions (Hagen, 1993). Since 1988, the air temperature has increased suggesting even larger difference between actual ELA and estimated ELA. In this context I would argue that Bogerbreen is less sensitive beyond the present climate, since the present day ELA has passed the elevation where Bogerbreen has its bulk of the area.

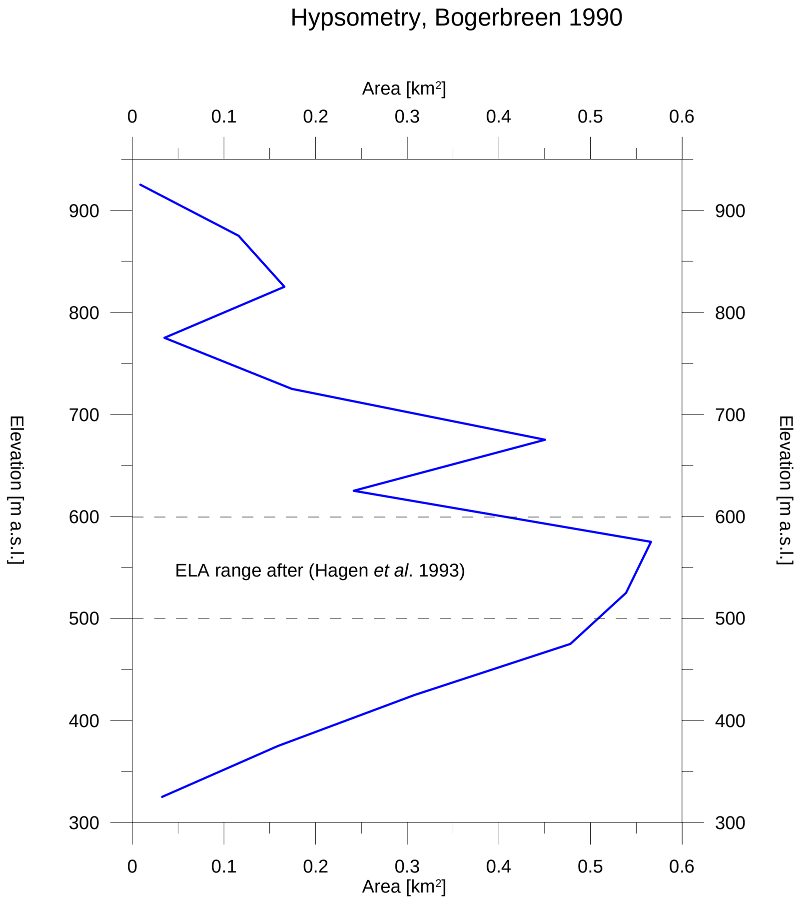

Figure 5.17 Hypsometry of Bogerbreens obtained from photogrammetrical survey of the 1990 summer surface. Indicated as the dashed lines is the zero balance ELA (Hagen, 1993).

Unfortunately the fragmentary mass balance record on Bogerbreen does not allow any detailed trend analysis (fig. 5.18). Earlier mass balance analysis of glaciers in Svalbard showed that there is no negative trend in the mass balance data on Svalbard despite the slight increase in summer temperatures and thus the accumulative sum of positive degree days (Dowdeswell *et al.*, 1997; Hagen and Liestøl, 1990). More recent publications indicate that there is a trend towards more negative balances observed during 1995 and 2003 (Ziaja, 2005; Bamber *et al.*, 2005). By comparison, the net balance mean over the period 1975 – 1985 (0.43 m water equivalent) with the mean over the period 1990 – 2004 (-0.65 m water equivalent) at Bogerbreen seems to indicate also increased negative mass balances due the climate introduced warming. Those results are supported by observations from Ziaja (2005). Ziaja (2005) concludes that on Grumantbreen, Håbergbreen and Dryadbreen, neighbouring glaciers to Bogerbreen, show signs of increased recession due to slight warming during the1990th (see also fig. 2.2).

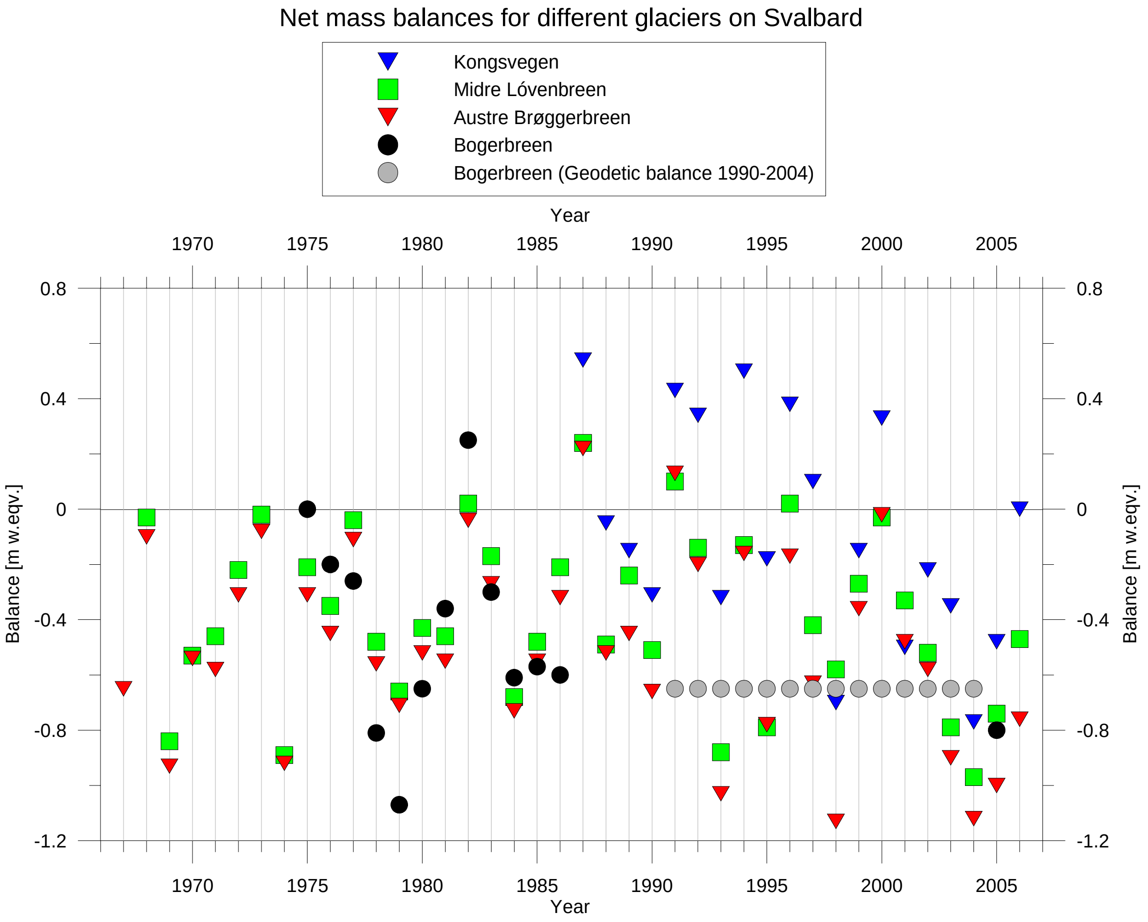

Figure 5.18. Total net mass balances for selected glaciers on Svalbard. Note that for Bogerbreen, the period between 1990/91 to 2003/04, is represented by the mean net balance for the period obtained by using the geodetic method. For all other glaciers presented here, balances have been measured using the direct glaciological method.

Only few glaciers on Svalbard were analyzed regarding their response to climate change (De Woul and Hock, 2005). It is very likely that Bogerbreen will decrease its volume further even without additional climate forcing (fig. 5.19). Konig *et al.* (2002) predicted for nearby glaciers that they would either vanish or only small areas of ice will remain at the highest elevation. I would infer a similar picture of Bogerbreen. A slight increase in precipitation has been observed on Svalbard during the last few decades with a simultaneous increase of positive degree days (Førland *et al.*, 1997; Hagen and Liestøl, 1990). Enhanced winter precipitation would possibly dampen the effect of increased summer melt but most likely not balance it out. Hagen and Liestøl (1990) concludes that on Svalbard an increase of mean summer temperature by 1 °C need to be compensated by 50 % increase of winter precipitation in order to maintain the same glacier mass balance. These results were obtained from Midre Lovénbreen and Austre Brøggerbreen. Compared to those, on Bogerbreen the control of net mass balance by summer mass balance is even more dominant. Therefore it is very likely, that even if present trends in winter precipitation continue, Bogerbreen will vanish or remain as a rather small cirque glacier.

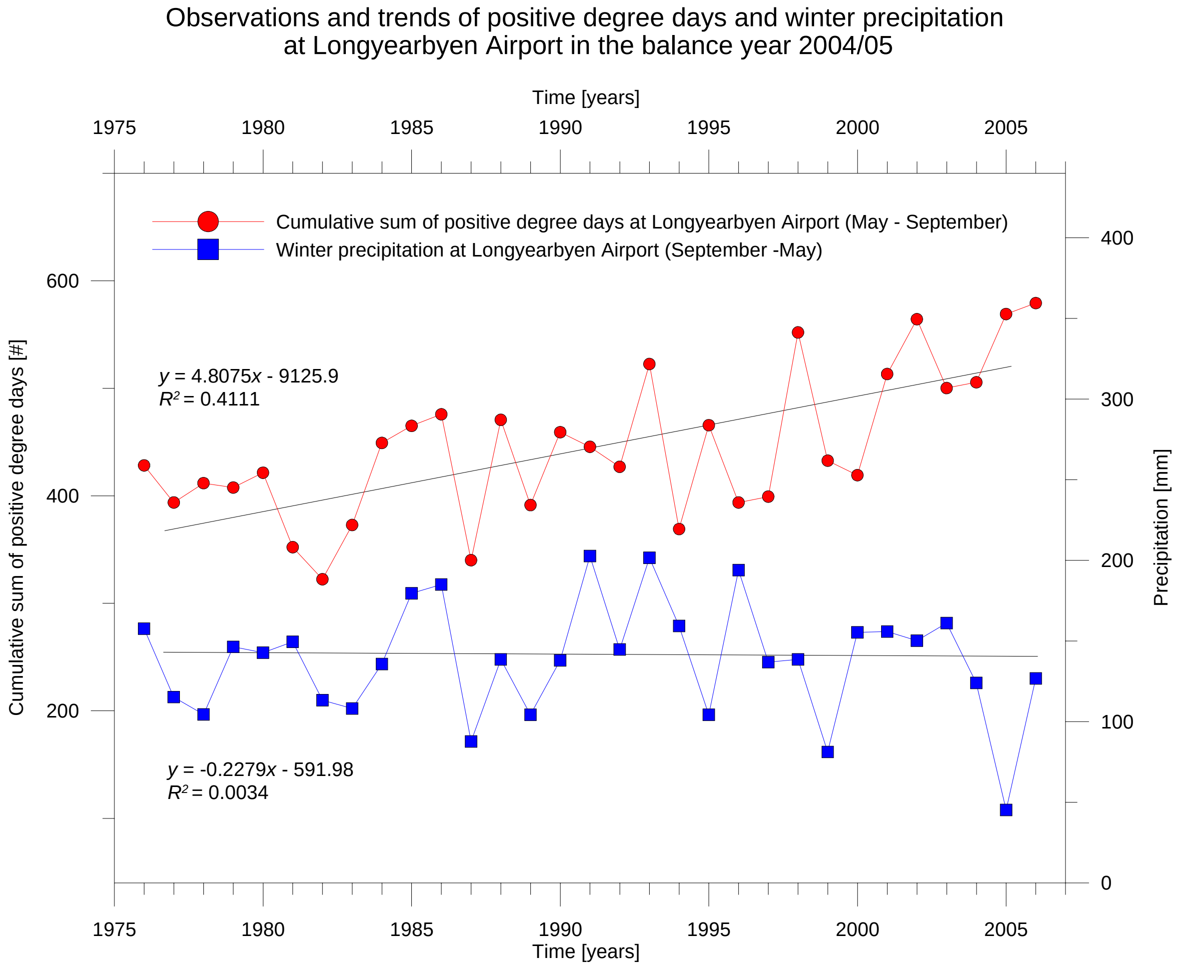

Figure 5.19 Meteorological observation at Longyearbyen Airport. The upper curve represents the cumulative sum of positive degree days with a linear trend fitted. The lower curve shows the cumulative sum of winter precipitation.

As outlined in the previous chapters, Bogerbreens net mass balance is dependent on various factors. It is difficult to predict the future behaviour by only considering for example precipitation and a proxy for summer ablation, such as positive degree days. Hodgkins *et al.* (2005) outlines in their studies how important the spatial distribution of winter accumulation is for the summer mass balance and consequently net mass balance. A change of the dominant wind direction, reduced wind speeds during winter are just two factors, which are likely to significantly alter Bogerbreens mass balance without seeing any change in the above mentioned meteorological observations. It seems to be almost impossible to distinguish important processes influencing mass balance on Bogerbreen by just looking at total mass balance values as they are obtained as a common routine on present glaciers. This discussion above should animate further studies to answer key question which still remained unanswered on Bogerbreen and on glaciers elsewhere.

6. Conclusion

6.1 Winter balance

- Bogerbreens winter balance 2004/05 was calculated to be 0.55 ± 0.08 m water equivalent. This is very similar to findings on Austre Brøggerbreen, Midre Lovénbreen, Kongsvegen and Linnébreen.

- In the balance year 2004/05, winter snow precipitation and most likely a significant amount of redistributed snow deposited onto the glacier, formed the largest contribution to the winter mass balance. Accumulation due to superimposed ice and avalanche deposits were insignificant.

- At the end of the winter season 2004/05, the spatial distribution of the winter mass balance was to a large degree controlled by the local wind pattern. The spatial distribution of the winter mass balance seems to have been consistent during the last few decades.

- Winter mass balance in 1975/76 to 1984/85 and in 2004/05 correlated only little with winter precipitation recorded at Longyearbyen Airport.

6.2 Summer balance

- Bogerbreens summer balance 2004/05 was calculated to be - 1.35 ± 0.24 m water equivalent. This is very similar to summer balances on Austre Brøggerbreen, Midre Lovénbreen. Kongsvegen shows less ablation while Linnébreen had enhanced ablation compared to Bogerbreen.

- The spatial distribution of winter mass balance at the beginning of the ablation season is likely to have a strong control on the spatial distribution of summer mass balance.

- It is likely that heat convection by air is the cause for the distinct summer balance gradient as a function of elevation. This gradient is further enhanced by albedo effects and influence of direct radiation. A combination of albedo effect and shading of direct radiation are possibly the cause for the lateral asymmetry in the spatial distribution of the summer mass balance.

- A reasonable good correlation exists between summer mass balance and cumulative positive degree days recorded at Longyearbyen Airport for the period1975/76 to 1984/85 and in 2004/05.

6.3 Net balance

- Bogerbreens net balance 2004/05 was negative, calculated to be - 0.80 ± 0.20 m water equivalent. This is very similar to net balances on Austre Brøggerbreen, Midre Lovénbreen. Kongsvegen shows less negative net mass balance while, Linnébreen had larger negative balance compared to Bogerbreen.

- The net mass balance of Bogerbreen is to a large degree controlled by the summer mass balance rather than the winter balance.

- It seems that the spatial pattern of net balance has been relatively persistent over the last few decades causing a major shift in the supraglacial drainage pattern and major implications for the local water supply in Longyearbyen.

- Bogerbreen has not been in balance with the climate in 2004/05, nor has it been during the last three decades.

- There is an indication suggesting that Bogerbreens negative net mass balance increased from the period 1975 to 1984 and 1990 to 2003 caused by climatic introduced warming. That is in accordance to recent observations on other Spitsbergen glaciers.

6.4 Uncertainties

- To determine winter mass balance, a pronounced spatial and most likely also temporal variation, demand a relatively dense network of measurements. The winter balance proofed to be insensitive to applying different inter- and extrapolation procedures for snow depth, as well as using different ways of applying density information. Thick ice lenses and careless snow density measurements can cause significant errors within the winter mass balance calculation.

- Lateral gradients of summer balance demands stakes located beside those along the centre line. The calculation of summer mass balance appeared to be sensitive to the method applied. Hand contouring proved to reduce the uncertainty by ± 0.13 m water equivalent or 10 % of the summer balance.

- It can not be excluded, that there is a discrepancy between mass balance results obtained for the period 1975/76 to 1984/85 and in 2004/05 on Bogerbreen.

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Appendix

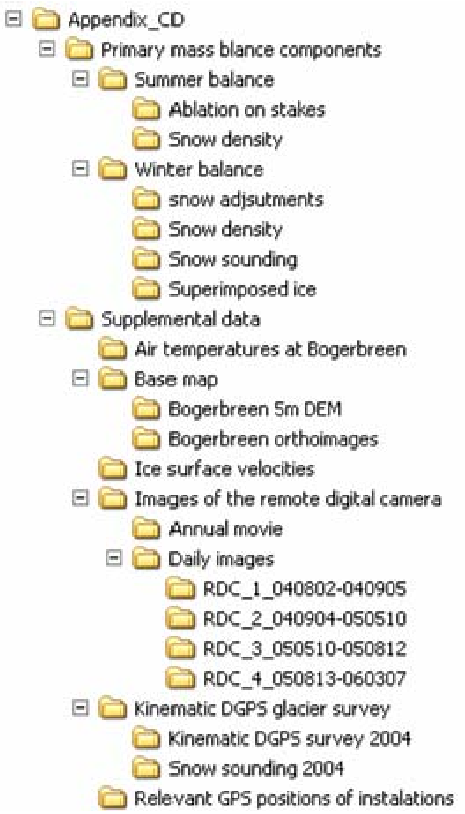
A CD-ROM, including raw data and results in a digital format, is attached in the cover page. The data organization on the media is visualized in figure A 1. The appendix CD contains:

Primary mass balance components 2004/05

- Summer ablation measurements (raw data)
- Snow density measurements (raw data)
- Initial snow survey measurements (raw data)
- Superimposed ice measurements (raw data)
- Snow survey adjustments (raw data)

Supplemental data

- Air temperature recordings on Bogerbreen (raw data)
- Daily digital images between August 2004 to March 2006 and a glacier movie (raw data and result)
- High resolution DEM (5 m) of Bogerbreen (result)
- Coordinates of the ground control points
- Orthophotographs of Bogerbreen (result)
- Stake positions surveyed 2004 and 2005 (result)
- Kinematic DGPS survey spring 2004 and snow survey 2004 (raw data)
- Relevant GPS positions on Bogerbreen for the period 2004 to 2006



- Appendix_CD
  - Primary mass blance components
    - Summer balance
      - Ablation on stakes
      - Snow density
    - Winter balance
      - snow adjsutments
      - Snow density
      - Snow sounding
      - Superimposed ice
  - Supplemental data
    - Air temperatures at Bogerbreen
    - Base map
      - Bogerbreen 5m DEM
      - Bogerbreen orthoimages
    - Ice surface velocities
    - Images of the remote digital camera
      - Annual movie
      - Daily images
        - RDC_1_040802-040905
        - RDC_2_040904-050510
        - RDC_3_050510-050812
        - RDC_4_050813-060307
    - Kinematic DGPS glacier survey
      - Kinematic DGPS survey 2004
      - Snow sounding 2004
    - Relevant GPS positions of instalations

Figure A 1 Data structure on the appendix CD-ROM