METEOROLOGICAL DIFFERENCES BETWEEN RABOTS GLACIÄR AND STORGLACIÄREN AND ITS IMPACT ON ABLATION

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Preface

This Master’s thesis is Pia Eriksson’s degree project in Physical Geography and Quaternary Geology at the Department of Physical Geography and Quaternary Geology, Stockholm University. The Master’s thesis comprises 60 credits (two terms of full-time studies).

Supervisor has been Peter Jansson at the Department of Physical Geography and Quaternary Geology, Stockholm University. Examiner has been Per Holmlund at the Department of Physical Geography and Quaternary Geology, Stockholm University.

The author is responsible for the contents of this thesis.

Stockholm, 12 December 2014

Lars-Ove Westerberg
Director of studies
Abstract

In the Kebnekaise Massif, Northern Sweden, the west facing glacier, Rabots glaciär, is losing volume at a significantly higher rate than east facing, Storglaciären. By analyzing data from automatic weather stations situated on the ablation area on the glaciers we investigated the effect of meteorological differences on ablation. There was a difference in micro-climate between Rabots glaciär and Storglaciären. Generally Storglaciären had slightly warmer and drier air, had less or a thinner cloud layer but more precipitation. On both glaciers a glacier wind is dominant but high wind velocities were common especially on Storglaciären indicating a larger influence from the synoptic system. There was a good correlation for temperature and vapor pressure between the glaciers that indicate that both glaciers are strongly affected by the synoptic system. The meteorological parameters have similar effect on the ablation on the glaciers. Temperature, vapor pressure and the turbulent heat fluxes are the only meteorological parameters that suggest a linear affect on ablation. Net shortwave radiation contribute with the greatest amount of energy for ablation but decreased in relative importance as the temperature increased. Shortwave radiation, sensible and latent heat contributed with a total 184 W m$^{-2}$ on Rabots glaciär and 222 W m$^{-2}$ on Storglaciären. Rabots glaciär seem to have a significantly greater relative importance of the turbulent heat fluxes than Storglaciären. Although the differences in micro-climate were not great, using the ablation for Storglaciären to estimate ablation on Rabots glaciär would over estimate the ablation with 0.5 m w.e.
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1. Introduction

A strategy to estimate regional change in glacier mass is to upscale data from well studied glaciers (Fountain et al., 1997). This benchmark approach has been used in most glaciated areas, often where information about change in mass is vital due to the importance of precise prediction of water runoff (e.g. Andreassen et al., 2005; Dyurgerov et al., 2006; Wagnon et al., 2007; Zhang et al., 2007; Huss et al., 2008; Fountain et al., 2009; Zemp et al., 2009). However, the benchmark glaciers are often chosen for logistic reasons and glaciers that are easy to access and relatively safe to work on are not necessarily the glaciers that are the most representative for the region (e.g. Dyurgerov et al., 2006; Fountain et al., 2009; Beusekom et al., 2010).

Fountain et al. (2009) studied how changes in a benchmark glacier in the North Cascade Range, USA, represented the changes in the region for the years 1993–2005. The benchmark glacier was larger and had a gentler slope than the average size and slope of the glaciers in the region. The study concluded that upscaling from the benchmark glacier led to an overestimation of mass change to the factor of three. However, the general pattern of the change in mass correlated well over the years indicating that the region was subjected to the same climate forcing and that the differences was caused by local topography.

In mountain terrain weather can change abruptly in space and time (Corby, 1954). Synoptic air flow, directly or indirectly, bring energy to the system through air temperature, air humidity, wind velocity and precipitation (Barry, 2008) and the air takes different paths depending on the size and shape of the barrier it encounters (Corby, 1954). It should, therefore, be possible that glaciers within the same region are not subject to the same climate forcing and that the mass balance of a glacier is affected by local topography due to glacier dynamics as well as climate forcing.

Several studies (e.g. Streten and Wendler, 1968; Hogg et al., 1982; Hay and Fitzharris, 1988; Braithwaite and Olesen, 1990; Munro, 1990; Hock and Holmgren, 1996; Konya et al., 2004; Hock and Holmgren, 2005; van de Wal et al., 2005; Andreassen et al., 2008; Giesen et al., 2008; Sicart et al., 2008; Giesen et al., 2009) have been done where more or less high temporal resolution meteorological data was used to compute the surface energy of individual glaciers. However, few studies have been done where the importance of the meteorological parameters are compared between adjacent glaciers. Wheler (2009) compared data from weather stations at two glaciers in the Donjek Range, St. Elias Mountains, US in order to evaluate different approaches to model melt and Giesen et al. (2009) compared 6 years of meteorology and surface energy data from Storbreen and Midtdalsbreen two glaciers 120 km apart in Norway.

In the Kebnekaise Massif, Northern Sweden, two adjacent, polythermal valley glaciers (Fig. 1.1) are known to be at different stages towards mass equilibrium (Stroeven and van de Wal, 1990). Stroeven and van de Wal (1990) and Brugger (2007) believe this to be caused by difference in glacier geometry and not by climate forcing. However, the possibility of a significant differences in micro-climate has not
yet been studied. Storglaciären and Rabots glaciär are part of the Tarfala mass balance programme but Storglaciären is considered a reference glacier (Holmlund and Jansson, 1999) and has therefore been the focus to more elaborate and detailed studies (e.g. Björnsson, 1981; Holmlund, 1987; Holmlund, 1988; Holmlund and Eriksson, 1989; Stroeven and van de Wal, 1990; Hock and Hooke, 1993; Hock, 1998; Jonsell et al., 2003; Jansson et al., 2007; Konya et al., 2007; Koblet et al., 2010). In July 2012 an automatic weather station, similar to the one already existing in the ablation area of Storglaciären, was installed in the ablation area of Rabots glaciär. This gave an opportunity to investigate how the micro-climate can vary between a benchmark glacier and a neighboring glacier.

By analyzing ablation data and data from the automatic weather stations our aim was to study the meteorological differences between Storglaciären and Rabots glaciär and how this affect ablation. This was done by focusing on following questions: *Do the micro-climate vary between the glaciers?*; *Do the individual meteorological parameters visibly affect ablation differently on the glaciers?*; *Do the magnitude of energy fluxes differ between the glaciers?*
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation

Figure 1.1. Photograph of a) Rabots glaciär and b) Storglaciären. Photo: Per Holmlund
2. Background

2.1. Mountain weather

2.1.1. Geographical control

The major factors that control mountain climate is latitude, continentality, altitude and topography (Barry, 2008). As a general rule, solar radiation and temperature decrease with increasing latitude (Marshak, 2001). As a consequence, the magnitude of the seasonal fluxes compared to the diurnal fluxes are inverted: at high latitudes the differences in temperature between summer and winter is significantly greater and at low latitudes the difference in temperature between day and night is greater (Troll, 1964). The seasonal and diurnal climatic rhythms also change with continentality due the heat capacity of water being significantly greater than terrestrial heat capacity (Driscoll and Fong, 1992). Closer to the ocean in the upwind direction difference in both seasonal and diurnal fluxes will decrease. Latitude and continentality are the greatest factors that will decide the major climate system for a specific mountain range (Barry, 2008). However, within the mountain range altitude and topography can cause great difference in weather over small distances (Corby, 1954).

![Figure 2.1](image)

*Figure 2.1*. The Earth’s energy budget. Incoming solar radiation (yellow arrows) reaches Earth’s atmosphere, where a part is absorbed by the atmosphere or reflected on clouds, aerosols or by molecular scattering. The remaining radiation is either absorbed or reflected at the surface. Heating of Earth’s surface and geothermal heat generates longwave radiation (red arrows). Most is absorbed by clouds and the atmosphere. Of the absorbed heat some is emitted from clouds and atmosphere into space and some is emitted back to the surface. (interpretation from Trenberth et al., 2009)
2.1.2. Altitude and topography

With increasing altitude, pressure and density are reduced and due to the saturation value controlled by temperature the vapor pressure is also likely to decrease (Barry, 1978). A relief over 600 m is believed to be the threshold where change in altitude begin to cause vertical differentiation of meteorological factors (Thompson, 1964). As unsaturated air rises the temperature decrease with 9.8 °Ckm⁻¹ (Barry, 2008). When saturated air cools as it rises it will undergo a condensation processes that release latent heat which lessen the cooling. Consequently the magnitude of the lapse rate of saturated air depend on the temperature of the air (Brunt, 1933). When the temperature is above 20 °C the lapse rate is approximately 5 °Ckm⁻¹ and for sub-zero conditions the available moisture is so small that the rate is greater due to significantly less release of latent heat and at −40 °C the lapse rate is almost equal to the unsaturated lapse rate (Barry, 2008).

When a global average of approximately 341 Wm⁻² incoming solar radiation reaches the atmosphere of the Earth (Trenberth et al., 2009) it either scatters back into space or gets absorbed by clouds, atmosphere and the surface of the Earth (Fig. 2.1). Globally, approximately 23 % of the incoming solar radiation scatters before reaching the ground (Trenberth et al., 2009) but the general low levels of aerosols in mountain areas together with a low amount of water vapor and a natural reduction of density, lessen the scattering at higher altitudes (Barry, 2008). Reduced scattering together with a naturally thinner cloud layer at high altitudes result in an increase of incoming solar radiation at increasing altitudes (Barry, 2008). However, the amount of energy that is absorbed is dependent of the albedo of the surface. A light surface will reflect a greater amount of radiation than a dark surface (Ångström, 1925). At high latitude Northern Scandinavia where snow covers the surface a great part of the year the mean energy absorbed by the surface is therefore likely to be significantly smaller than the global average of 161 Wm⁻² (Trenberth et al., 2009). In Kiruna, Northern Sweden the normal (mean values over the years 1961–1990 as defined by the World Meteorological Organization) mean net solar radiation is 92.5 Wm⁻² (Data obtained from SMHI, http://www.smhi.se/klimatdata/meteorologi/2.1240, 9 Sep., 2014).

Where the mid-latitude Westerlies prevails the wind speed will increase with height and exposed ridges and peaks are usually subject to even higher winds speeds due to limited friction (Barry, 2008). When a weather system reaches a barrier the potential energy within the system and the energy needed to pass the barrier determines how the system will respond (Barry, 2008). Topography therefore influences weather systems depending on the three dimensional size and direction of the system but also the three dimensional size and shape of the barrier. It is common that when air passes over a mountain it begins to flown in a wave motion on the lea side which causes turbulence (Corby, 1954).

Due to the physical effects of topography on air flow different types of fall winds occur down the lee slope of mountains (Barry, 2008). In mountain areas it is common that a rain shadow effect occurs on the lee side of mountains. When moist air rises on the up-wind side of the mountain and is cooled to the point of condensation it releases precipitation (Barry, 2008). The falling air on the lee side is now close to an unsaturated state which causes a greater heating and consequently a lowering of relative humidity (Brinkmann, 1971).

The thermal differentiations due to topography also causes patterns of air flow
motions (Barry, 2008). Most commonly these vertical or horizontal motions are caused by elevation differences in potential temperature and uneven heating and cooling of slopes. *Slope winds* (or more correctly, slope breezes) are divided into *katabatic* flow and *anabatic* flow (Barry, 2008). Katabatic flow are downslope gravity flow caused by surface cooling at night. At daytime the lower part of the slope heats faster and, due to buoyancy, flow upslope (Barry, 2008). Mountain and valley winds are caused by the same processes as the slope breezes but are greater in scale and velocity. The mountain wind flow down-valley at night and the valley wind up-valley during the day. The valley wind is approximately 1 km thick and mixes frequently with the slope breeze (Oerlemans, 2010) whereas the mountain wind is shallower and has lower velocities (Barry, 2008).

\[\text{Figure 2.2. Illustration of the acting air flow on a valley glacier. The geostrophic flow and large-scale boundary layer is relatively unaffected by the topography whereas the valley wind and the glacier wind is controlled by the topography (rework from Oerlemans, 2001).}\]

### 2.1.3. Permanent snow and ice

In summer, the most striking difference between permanent snow and ice and its surroundings is the significantly lower temperature and higher albedo of the snow and ice (Oerlemans, 2010). In mid- and high-latitude mountains the surface changes dramatically between the seasons but for permanent snow and ice the changes are significantly smaller (Oerlemans, 2010). However, even though the changes are relatively small the surface of a glacier changes constantly. The ablation area often evolve from a smooth snow cover to a rough ice surface with cryoconites and debris within a few months (Benn and Evans, 2010). The surface roughness affects the solar reflectance and consequently the amount of solar radiation that will be absorbed by the surface. Table 2.1 show the albedo of snow and ice from studies on glaciers around the world (Jonsell et al., 2003; Hock and Holmgren, 1996; Wallén, 10).
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1949; Andreassen et al., 2008; Giesen et al., 2008; Braithwaite, 1995; van de Wal et al., 2005; Escher-Vetter, 1985; Hannah et al., 2000; MacDougall, 2010; Konya and Matsumoto, 2010). The thresholds for snow- and ice- albedo differ but a general pattern is evident. Albedos for fresh snow is high (0.8–0.9), old or wet snow is lower (0.5–0.7) and the albedo for ice is low and extremely diverse (0.15–0.51).

Table 2.1. Albedo values for several studies around the world.

<table>
<thead>
<tr>
<th>Location (Reference)</th>
<th>Snow α</th>
<th>Ice α</th>
</tr>
</thead>
<tbody>
<tr>
<td>Storglaciären, Sweden (Jonsell et al., 2003)</td>
<td>0.54–0.93 [1]</td>
<td>0.22–0.51</td>
</tr>
<tr>
<td>Storglaciären, Sweden (Hock and Holmgren, 1996)</td>
<td>0.62–0.88 [1]</td>
<td>0.42</td>
</tr>
<tr>
<td>Storglaciären, Sweden (this study)</td>
<td>&gt; 0.6 [1]</td>
<td>&lt; 0.4</td>
</tr>
<tr>
<td>Rabots glaciär, Sweden (this study)</td>
<td>&gt; 0.7 [1]</td>
<td>&lt; 0.5</td>
</tr>
<tr>
<td>Kårsaglaciären, Sweden (Wallén, 1949)</td>
<td>0.59–0.81 [1]</td>
<td>0.36</td>
</tr>
<tr>
<td>Storbreen, Norway (Andreassen et al., 2008)</td>
<td>0.9 [2]</td>
<td>0.3</td>
</tr>
<tr>
<td>Midtdalsbreen, Norway (Giesen et al., 2008)</td>
<td>0.7 [3]</td>
<td>0.31</td>
</tr>
<tr>
<td>Nordbogletscher and Qamanarsûp sermia, Greenland (Braithwaite, 1995)</td>
<td>0.7 [4]</td>
<td>0.3</td>
</tr>
<tr>
<td>Kangerlussuaq transect, Greenland (van de Wal et al., 2005)</td>
<td>0.70 [3]</td>
<td>0.55</td>
</tr>
<tr>
<td>Vernagtferner, Austria (Escher-Vetter, 1985)</td>
<td>0.8 [2]</td>
<td>0.4</td>
</tr>
<tr>
<td>Taillon Glacier, France (Hannah et al., 2000)</td>
<td>0.58 [4]</td>
<td>0.28</td>
</tr>
<tr>
<td>The Donjek Range, USA (MacDougall, 2010)</td>
<td>0.66–0.90 [1]</td>
<td>&gt; 0.15</td>
</tr>
<tr>
<td>Glaciar Exploradores, Chile (Konya and Matsumoto, 2010)</td>
<td>–</td>
<td>0.19 &amp; 0.37</td>
</tr>
</tbody>
</table>


The atmosphere feeds the snow and ice surface with energy causing ablation processes but the snow and ice itself influence the atmosphere by its presence (Hock, 2005). The surface can never be above the melting point and consequently during the melt season a large temperature gradient occur close to the surface (Oerlemans, 2010). In sub-zero conditions the snow can be warmer than the surrounding air creating a vertical temperature gradient in the opposite direction. However, sublimation occurs constantly when the snow or ice crystals are directly or indirectly exposed to a medium containing higher energy. Therefore the process enhances with high radiation, high temperature or high wind velocity when the air expose the crystals either by transport or by penetrating the surface through pores into the ice or snowpack (Schmidt and Gluns, 1992). The sublimation process absorbs $2.83 \times 10^6 \text{Jkg}^{-1}$ latent heat, which is the summarized latent heat absorption of melting and evaporation (Strasser et al., 2008). Consequently the sublimation process cools the air as much as melting and evaporation combined which can create a vertical temperature towards the surface. The stratification of temperature on a melting glacier is relatively stable and suppress turbulence. During the melt season the glacier surface will be colder than the surrounding and the stratified temperature profile causes katabatic flow down the glacier. This flow, also known as the glacier wind, is a shallow wind (approx. 20 m thick) not higher than 5 m s$^{-1}$ (Oerlemans and Grisogono, 2002).

Figure 2.2 illustrates the interaction between the valley wind and the glacier wind. The large scale processes, the geostrophic flow, is relatively unaffected by the topography whereas the underlying boundary layer is slightly tilted due to drag from the topography. Underneath is the valley wind system flowing up valley and the glacier wind flowing down valley. In front of the glacier the two systems meet forcing
the lighter warm valley wind on top of the heavier glacier wind. This interaction between the valley wind and the glacier wind is the glaciers main source for heat exchange (Oerlemans, 2010).

2.2. Ablation parameters

The meteorological factors and the physical properties of the glacier determine the surface energy balance (Hock, 2005). Figure 2.3 illustrates the most important processes that determine the energy flux of a glacier. The largest exchange come from the radiation fluxes. Considering the variation of the albedo of glacier surfaces (Table 2.1) the rate of solar radiation that will be reflected is dependent on the surface properties. When covered with debris significantly more radiation is absorbed by the surface. The solar radiation can penetrate approximately 10 m of snow and 1 m of ice but only 1% to 2% penetrates into the pack due to efficient absorption of energy in the upmost cm of the snow pack (Hock, 2005). Even though the effect is small this process can be important considering it can result in internal melt in sub zero conditions. The incoming longwave radiation varies and is all absorbed by the surface whereas the outgoing longwave radiation is high and relatively constant, leading to low net longwave values (Oerlemans and Grisogono, 2002). Often the net longwave radiation is negative but in warm and humid overcast conditions it can be positive (Oerlemans, 2001). The shortwave and longwave radiation has a reverse response to clear and overcast conditions. The presence of clouds will lower the incoming shortwave radiation but heighten the longwave radiation (Oerlemans, 2010). The magnitude of the response is, however, strongly connected to the properties of the clouds and the surface albedo, indicating that both an increase and decrease in net radiation is possible (Oerlemans, 2001). In increasing overcast conditions Giesen et al. (2009) observed an increase in net radiation over a snow surface and a decrease over an ice surface.

The magnitude of sensible and latent heat (turbulent heat fluxes) are primarily affected by the air temperature and relative humidity, respectively. In summer, the magnitude of turbulent exchange of heat (sensible heat flux) and moisture (latent heat flux) is greater than in winter but as a result of the low incoming solar radiation in winter the relative importance of the turbulent heat fluxes are higher.
**Table 2.2.** Mean values of temperature (°C), wind speed (m s\(^{-1}\)) and energy fluxes (W m\(^{-2}\)) from energy balance studies around the globe. Numbers in brackets are the relative contribution to the surface energy balance. Values are rounded to the nearest integer. For variable names, see subsection 4.3.3.

<table>
<thead>
<tr>
<th>Location (Reference)</th>
<th>Period</th>
<th>(\bar{T})</th>
<th>(\bar{s}_w)</th>
<th>(\bar{I}\downarrow)</th>
<th>(I + L)</th>
<th>(\bar{H}_s)</th>
<th>(\bar{H}_l)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Storglaciären, Sweden (Hock and Holmgren, 1996)</td>
<td>19 Jul.–20 Aug., 1994</td>
<td>5.4</td>
<td>2.5</td>
<td>73 (66)</td>
<td>33 (30)</td>
<td>5 (5)</td>
<td></td>
</tr>
<tr>
<td>Storglaciären, Sweden (Hock and Holmgren, 2005)</td>
<td>7 Jun.–17 Sep., 1993</td>
<td>–</td>
<td>–</td>
<td>147 18 (38)</td>
<td>20 (43)</td>
<td>8 (17)</td>
<td></td>
</tr>
<tr>
<td>Storglaciären, Sweden (Hock and Holmgren, 2005)</td>
<td>5 Jun.–6 Sep., 1994</td>
<td>–</td>
<td>–</td>
<td>169 49 (58)</td>
<td>36 (42)</td>
<td>0 (0)</td>
<td></td>
</tr>
<tr>
<td>Storglaciären, Sweden (Sicart et al., 2008)</td>
<td>9 Jul.-2 Sep., 2000</td>
<td>5.4</td>
<td>–</td>
<td>133 59 (55)</td>
<td>35 (32)</td>
<td>14 (13)</td>
<td></td>
</tr>
<tr>
<td>Storglaciären, Sweden (this study)</td>
<td>16 May–5 Sep., 2013</td>
<td>5.6</td>
<td>3.0</td>
<td>181 82 (67)</td>
<td>32 (26)</td>
<td>9 (7)</td>
<td></td>
</tr>
<tr>
<td>Rabots glaciär, Sweden (this study)</td>
<td>16 May–5 Sep., 2013</td>
<td>4.8</td>
<td>2.5</td>
<td>149 –</td>
<td>28 7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Storbreen, Norway (Andreassen et al., 2008)</td>
<td>1 Jun.–10 Sep., 2002–06</td>
<td>5.3</td>
<td>3.2</td>
<td>185 86 (76)</td>
<td>20 (18)</td>
<td>9 (8)</td>
<td></td>
</tr>
<tr>
<td>Midtdalsbreen, Norway (Giesen et al., 2008)</td>
<td>Melt period, 2001–05</td>
<td>–</td>
<td>–</td>
<td>242 101 (67)</td>
<td>37 (25)</td>
<td>15 (10)</td>
<td></td>
</tr>
<tr>
<td>Midtdalsbreen, Norway (Giesen et al., 2009)</td>
<td>Melt period, 2001–06</td>
<td>5.3</td>
<td>6.0</td>
<td>242 104 (66)</td>
<td>39 (25)</td>
<td>16 (10)</td>
<td></td>
</tr>
<tr>
<td>Storbreen, Norway (Giesen et al., 2009)</td>
<td>Melt period, 2001–06</td>
<td>4.9</td>
<td>3.3</td>
<td>220 90 (77)</td>
<td>20 (17)</td>
<td>9 (8)</td>
<td></td>
</tr>
<tr>
<td>Nordbogletscher, Greenland (Braithwaite and Olesen, 1990)</td>
<td>Jun.–Aug., 1979–83</td>
<td>–</td>
<td>–</td>
<td>80 (73)</td>
<td>32 (29)</td>
<td>-2 (-2)</td>
<td></td>
</tr>
<tr>
<td>Qamanârâssûp sermia, Greenland (Braithwaite and Olesen, 1990)</td>
<td>Jun.–Aug., 1979–83</td>
<td>–</td>
<td>–</td>
<td>107 (67)</td>
<td>61 (38)</td>
<td>-8 (-5)</td>
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<td>K-transect, Greenland (van de Wal et al., 2005)</td>
<td>Jun.–Aug.1998–02</td>
<td>0.3</td>
<td>4.6</td>
<td>260 61</td>
<td>–</td>
<td>–</td>
<td></td>
</tr>
<tr>
<td>Peyto Glacier (Ice), Canada (Munro, 1990)</td>
<td>17 Jun.–6 Jul.,1988</td>
<td>5.7</td>
<td>3.9</td>
<td>202 108 (65)</td>
<td>57 (34)</td>
<td>2 (1)</td>
<td></td>
</tr>
<tr>
<td>Peyto Glacier (Snow), Canada (Munro, 1990)</td>
<td>21 Jun.–5 Jul.,1988</td>
<td>3.7</td>
<td>3.0</td>
<td>199 39 (51)</td>
<td>32 (42)</td>
<td>5 (7)</td>
<td></td>
</tr>
<tr>
<td>Worthington Glacier, USA (Stretten and Wendler, 1968)</td>
<td>16 Jul.–1 Aug., 1967</td>
<td>9.6</td>
<td>2.1</td>
<td>127 (51)</td>
<td>68 (29)</td>
<td>47 (20)</td>
<td></td>
</tr>
<tr>
<td>St Sorlin, France Alps (Sicart et al., 2008)</td>
<td>9 Jul.–27 Aug., 2006</td>
<td>5.4</td>
<td>–</td>
<td>233 127 (84)</td>
<td>33 (22)</td>
<td>-8 (-5)</td>
<td></td>
</tr>
<tr>
<td>Hodges Glacier, South Georgia (Hogg et al., 1982)</td>
<td>1 Nov.–4 Apr., 1973–74</td>
<td>5.6</td>
<td>3.9</td>
<td>284 47 (55)</td>
<td>41 (48)</td>
<td>-2 (-3)</td>
<td></td>
</tr>
<tr>
<td>Kuryo Glacier, Russia (Konya et al., 2004)</td>
<td>7 Aug.–12 Sep., 2000</td>
<td>7.6</td>
<td>2.4</td>
<td>43 (33)</td>
<td>59 (44)</td>
<td>31 (23)</td>
<td></td>
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<tr>
<td>Ivory Glacier, New Zealand (Hay and Fitzharris, 1988)</td>
<td>6 Jan.–14 Feb., 1972</td>
<td>–</td>
<td>–</td>
<td>278 68 (54)</td>
<td>44 (35)</td>
<td>15 (12)</td>
<td></td>
</tr>
<tr>
<td>Zongo, Bolivia (Sicart et al., 2008)</td>
<td>1 Nov.–21 Dec., 1999</td>
<td>0.2</td>
<td>–</td>
<td>214 64 (103)</td>
<td>15 (24)</td>
<td>-17 (-27)</td>
<td></td>
</tr>
</tbody>
</table>
Table 2.3 show the magnitude of the average incoming radiation, sensible heat flux and latent heat flux at different time span within the melt season around the world. The mean sensible heat flux ranges from 15 W m$^{-2}$ in Zongo, Bolivia where the mean temperature is 0.2°C (Sicart et al., 2008) to 68 W m$^{-2}$ on Worthington Glacier, USA where the mean temperature is 9.6°C (Streten and Wendler, 1968). The mean latent heat flux ranges from $-17$ W m$^{-2}$ to $47$ W m$^{-2}$ on the same glaciers as above (Sicart et al., 2008; Streten and Wendler, 1968). The range of incoming shortwave radiation is $133$ W m$^{-2}$ to $260$ W m$^{-2}$ and is strongly dependent on latitude, altitude and weather.

Precipitation will add or remove heat depending of the temperature of the precipitation compared to the surface. These fluxes are however small and only make up a few percent of the energy balance even if the amount of rain is extreme (Giesen et al., 2009). Due to the low saturation vapor pressure at a melting surface (611 Pa) there is a vapor pressure gradient that will be towards or from the surface depending on the humidity of the air. Condensation is an important source of energy and will in high humidity conditions heat the surface. Evaporation on the other hand consume great amounts of energy which consequently cools the surface (Oerlemans, 2001).

Except flow of meltwater which is a latent heat process the internal energy fluxes (ground heat fluxes) shown in figure 2.3 is significantly smaller than the fluxes interacting between the atmosphere and the glacier surface. Molecular conduction (diffusion) and convection by air motion that transports heat and moisture are small fluxes that are mostly important for the metamorphism of snow crystals. In a temperate glacier the ground heat flux is negligible but in a polythermal or polar glacier it might have a small negative contribution to the energy balance. Hock and Holmgren (1996) reported a contribution of $-3\%$ on Storglaciären, Sweden and Giesen et al. (2009) reported $-2\%$ on both Storbreen and Midtdalsbreen, Norway.
3. Study area

Kebnekaise massif

The studied glaciers are situated in the Kebnekaise massif (67.9° N, 18.5° E) approximately 70 km west of Kiruna, Northern Sweden. The massif consists of a handful glaciers separated by 2000 m a.s.l. peaks (Fig. 3.1). The climate in the region is considered continental and the prevailing wind is westerly (Holmlund and Jansson, 1999). Figure 3.2 show the normal (mean values over the years 1961–1990 as defined by the World Meteorological Organization) monthly values for temperature, net shortwave radiation and total precipitation for Kiruna (Data obtained from SMHI, http://www.smhi.se/klimatdata/meteorologi/2.1240, 9 Sep., 2014). It also presents the monthly mean for 2013 and the monthly highest and lowest values ever recorded at the station. Normally the temperature peaks in July (12.8°C) and is lowest in January (−15.6°C). The net shortwave radiation peaks in June (219.3 W m⁻²), disappears entirely in the middle of November and returns in January. The precip-

![Figure 3.1. The glaciers of the Kebnekaise massif in Northern Sweden. Separating the glaciers is a steep ridge containing the two highest peaks in Sweden. Lakes and larger streams in blue and the red crosses point out the location of weather stations.](image-url)
Iteration is relatively evenly distributed over the year with a slight peak in summer (82 mm in July).

![Figure 3.2](image)

*Figure 3.2. Normal values for weather data from Kiruna, Northern Sweden. a) Mean temperature, b) net shortwave radiation, c) total precipitation (Data obtained from SMHI, http://www.smhi.se/klimatdata/meteorologi/2.1240, 9 Sep., 2014).*

The glaciers in the Kebnekaise massif have been studied for decades and the earliest records date from over a century ago (Holmlund et al., 1996). Storglaciären (Fig. 1.1a) and Rabots glaciär (Fig. 1.1b) have the the longest mass balance series (Fig. 3.3) in Sweden.

**Rabots glaciär**

Rabots glaciär is a small polythermal valley glacier (Fig. 1.1) (Schytt, 1959) in the Kebnekaise massif. It has an area of $3.7 \text{ km}^2$ (Brugger et al., 2005) and a mean thickness of approximately 85 m and a maximum thickness of 175 m (Brugger et al., 2005). It consists of a large ablation area in a north-east to south-west direction with slope angles ranging from 4° to 12° (Stroeven and van de Wal, 1990) and three cirques that make up the accumulation area (Fig. 3.4). Surrounding topography is relatively high and steep but the bottom topography is believed to be flat, lacking overdeepenings (Björnsson, 1981).

Rabots glaciär reached its Holocene maximum extent around 1916 (Karlén, 1973) and started to retreat in the late 1920s. The glacier is believed to not have reached equilibrium after the temperature increase following the little ice age. Brugger et al. (2005) and Brugger (2007) studied ice margin retreats and modeled response of Rabots glaciär to the temperature increase and concluded that the glacier has twice as long response time as the neighboring glacier, Storglaciären. This is believed to
be caused by glacier geometry and not difference in meteorology or hydrology. A similar conclusion was drawn by (Stroeven and van de Wal, 1990) who compared mass balance and flow between Rabots glaciär and Storglaciären. They concluded that the mass balance pattern was comparable and that the slightly more ablation at Rabots glaciär was due to the glacier being in a state of non equilibrium.

The mass at Rabots glaciär have been measured since the mass balance year 1981/82 (Holmlund and Jansson, 1999) and since then it has decreased with just over 12 m water equivalent (Fig. 3.3). Even though Rabots glaciar have been monitored for a relatively long period of time only a few studies have been concentrated on the glacier (Stroeven and van de Wal, 1990; Brugger et al., 2005; Brugger, 2007).

Figure 3.4. Schematic over Rabots glaciär. The red cross is the position of the weather station and the dotted line is the approximate snow line in August 2013 mapped from aerial photograph obtained from Lantmäteriet, http://www.lantmateriet.se, 24 Mar., 2014.
**Storglaciären**

Storglaciären is situated on the opposite side of the ridge from Rabots glaciär (Fig. 3.1). It is Sweden’s most studied glacier and has the longest record of mass balance in the world (1945/46–present). It is a small polythermal valley glacier (Schytt, 1959), with an area slightly smaller than Rabots glaciär (3.1 km$^2$ according to Brugger et al., 2005). It has a mean thickness of 100 m and a maximum thickness of 250 m (Björnsson, 1981). The glacier stretches from west to east and has a large, relatively flat ablation area and two steeper cirques that make up the accumulation area (Fig. 3.5). In contrast to the flat bottom topography of Rabots glaciär, the steep and rough surrounding topography continues underneath the Storglaciären where three overdeepenings can be found (Björnsson, 1981). The Holocene maxi-

![Figure 3.5. Schematic over Storglaciären. The red cross is the position of the weather station and the dotted line is the approximate snow line in August 2013 mapped from aerial photograph obtained from Lantmäteriet, http://www.lantmateriet.se, 24 Mar., 2014.](image)

...mum extent of Storglaciären occurred around 1916 and the glacier started retreating in the late-1920s (Karlén, 1973). In the mid-1980s the glacier terminus started to stabilize and increased slightly in volume. This is believed to be caused by an increased maritime climate forcing which brought more winter precipitation (Stroeven and van de Wal, 1990). Since then the glacier is believed to be close to equilibrium (Holmlund, 1988).

The study of the mass balance of Storglaciären started in 1945 (Fig. 3.3). Since then the glacier have been the focus on numerous studies concerning mass balance measurements and reconstructions, modelling, hydrology, thermal properties etc. (e.g. Björnsson, 1981; Holmlund, 1987; Holmlund, 1988; Holmlund and Eriksson, 1989; Stroeven and van de Wal, 1990; Hock and Hooke, 1993; Hock, 1998; Jonsell et al., 2003; Jansson et al., 2007; Konya et al., 2007; Koblet et al., 2010).
4. Methodology

4.1. Data collection

From the middle of April to early September continuous meteorological and surface melt data was recorded by automatic weather stations situated on the ablation area on Storglaciären (67.9030° N, 18.5726° S, 1355 m a.s.l.) and Rabots glaciär (67.9112° N, 18.4925° S, 1373 m a.s.l.) (Fig. 3.1). The weather stations measured; wind speed, wind direction, temperature, relative humidity, precipitation, radiation (incoming and reflected shortwave radiation on Rabots glaciär and incoming and outgoing long- and shortwave radiation on Storglaciären). Coupled to the data logger was also a sonic ranger that continuously measured the distance to the surface.

4.1.1. Automatic weather stations

The Automatic Weathers Stations consisted of several separate instruments (Fig. 4.1) with sensors (Table 4.1) measuring different meteorological parameters. The instruments were connected to a logger that recorded data at a time interval according to a logger program (a subsection A.1.2). Except for different radiation sensors the stations on Rabots glaciär and Storglaciären were identical. Figure 4.1g shows the setup of the station on Storglaciären. The instruments are mounted directly to a 0.05 m diameter mast or to a 0.04 m diameter crossarm mounted on the mast. Temperature and humidity are measured at three heights, 0.5 m, 1 m and 2 m above the glacier surface. On the crossarm are instruments measuring rainfall (Fig. 4.1b), wind speed and wind direction (Fig. 4.1c) and radiation (Fig. 4.1d or e). Mounted on the mast are also the logger box (i in figure 4.1g), containing a datalogger and a 12V battery, and a solar panel (h in figure 4.1g) for continuous charging of the battery. The automatic weather stations are placed on the glacier surface at the end of winter so the height of the instruments in relationship to the surface will be as constant as possible. The station on Rabots glaciär was installed 25 March, 2013 and dismantled 6 September, 2013. The sonic ranger were installed 14 April, 2013. At Storglaciären the station and sonic ranger were installed 17 April, 2013 and dismantled 9 September, 2013.

Campbell HC2S3, Temperature and relative humidity probe

The HC2S3, Temperature and Relative Humidity Probe have two different sensors (Table 4.1). The 100 Ω PRT sensor measured the temperature and the ROTRONIC Hygromer® IN1 capacitive sensor measures the relative humidity. The HC2S3 has a polyethylene filter that protects the sensor from fine aerosols and minimize the water absorption and retention (CS, 2012a).

The range of the temperature sensor is −40°C to 60°C and it measures with the accuracy of ±0.1°C to 0.3°C depending on the temperature. The relative humidity
Figure 4.1. The Automatic weather station instruments: a) Temperature and Relative Humidity Probe where 1 is the radiation shield and 2 the sensors; b) Wind Monitor; c) Tipping Bucket Raingauge; d) Pyranometer; e) Net Radiation Sensor where 3 is the pyranometer and 4 is the pyrgeometer; f) Sonic Ranging Sensor; g) The setup of the station on Storglaciären; h) Solar Panel; i) Logger Box.
### Table 4.1. Name, range and accuracy of the Campbell instrument and sensor used in the study.

<table>
<thead>
<tr>
<th>Instrument, Sensor</th>
<th>Measurement (unit)</th>
<th>Range</th>
<th>Accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>HC2S3, Temperature and Relative Humidity Probe, PT100 RTD, IEC 751 1/3 Class B, with calibrated signal conditioning</td>
<td>Temperature (°C)</td>
<td>-40–60</td>
<td>±0.1[^1] ±0.2[^2] ±0.3[^3]</td>
</tr>
<tr>
<td>HC2S3, Temperature and Relative Humidity Probe, ROTTORIC Hygrometer® IN1</td>
<td>Relative humidity (%)</td>
<td>0–100</td>
<td>±0.8[^1] ±1.3[^2] ±3.3[^3]</td>
</tr>
<tr>
<td>CS300, Pyranometer</td>
<td>Shortwave radiation (300–1100 nm)</td>
<td>0–2000</td>
<td>±5%[^8]</td>
</tr>
<tr>
<td>NR01, 4-component Net Radiation Sensor, SR01 Pyranometers</td>
<td>Shortwave radiation (305–2800 nm)</td>
<td>0–2000</td>
<td>±10%[^8]</td>
</tr>
<tr>
<td>NR01, 4-component Net Radiation Sensor, IR01 Pyrgeometers</td>
<td>Longwave radiation (4500–50 000 nm)</td>
<td>0–1000</td>
<td>±10%[^8]</td>
</tr>
<tr>
<td>05103, Wind Monitor, 180 mm 4-blade helicoid propeller</td>
<td>Wind speed (m s⁻¹)</td>
<td>0–100</td>
<td>±0.3[^6] ±1.0[^7]</td>
</tr>
<tr>
<td>05103, Wind Monitor, Balanced vane, 380 mm turning radius</td>
<td>Wind direction (°)</td>
<td>0–360</td>
<td>±3</td>
</tr>
<tr>
<td>52202, Tipping Bucket Raingauge</td>
<td>Precipitation (mm)</td>
<td>—</td>
<td>±2%[^4] ±3%[^5]</td>
</tr>
<tr>
<td>SR50A, Sonic Ranging Sensor</td>
<td>Distance to ground (m)</td>
<td>0.5–10</td>
<td>±0.01%[^9]</td>
</tr>
</tbody>
</table>

Accuracy at/for:[^1] at 23 °C,[^2] at −10 °C,[^3] at −40 °C,[^4] for 0–25 mm h⁻¹,[^5] for 25–50 mm h⁻¹,[^6] for 1–60 m s⁻¹,[^7] for 60–100 m s⁻¹,[^8] of total daily radiation,[^9] ± 0.4% when the distance is > 2.5 m

Sensor range is 0% to 100% and has an accuracy of ±0.8% to 3.3% depending on the temperature. Both sensors recorded transients or values based on 10 s or 60 s averages (Table A.1).

Three HC2S3 were each placed within a radiation shield (41003-5) (Fig. 4.1) and mounted on the 0.05 m diameter mast at 0.5 m, 1 m and 2 m above the glacier surface (Fig. 4.1g).

**Campbell 52202, Tipping bucket raingauge**

The 52202, *Tipping Bucket Raingauge* measures the precipitation (Table 4.1). The precipitation falls into a 200 cm catchment area and gets funneled into a bucket that tips when the bucket fills up to 0.1 mm (CS, 2010). The movement sets off a reed switch that sends a pulse to the datalogger which records the amount of pulses from the reed switch.

The tipping bucket has an accuracy of ±2% to 3% depending on intensity of the rain. It recorded the sum of precipitation every 15 min (Table A.1).

The 52202 was mounted in the center of the crossarm (Fig. 4.1g).

**Campbell 05103, Wind monitor**

The 05103, *Wind Monitor* has two sensors (Table 4.1). The 180 mm diameter 4-blade helicoid propeller molded from polypropylene measures wind speed and the Balanced vane with a 380 mm turning radius measures the wind direction (CS, 2012b). Rotation of the propeller produces an alternating current sine wave with a frequency
directly proportional to wind speed. The balanced vane uses a potentiometer to measure wind direction. The output is proportional to the azimuth angle which has to be manually measured to acquire correct cardinal point.

The range of the wind speed sensor is $1 \text{ m s}^{-1}$ to $100 \text{ m s}^{-1}$ and the sensor measures with an accuracy of $\pm 0.3 \text{ m s}^{-1}$ to $1 \text{ m s}^{-1}$ depending on wind speed. Every 15 min average, maximum, minimum and standard deviation is recorded (Table A.1).

The range of the wind direction sensor is $0^\circ$ to $360^\circ$ and the sensor has an accuracy of $\pm 3\%$. The sensor is read every 10 s and records an averages and standard deviation every 15 min (Table A.1).

On both glacier the 05103 was mounted on the north facing end of the crossarm (Fig. 4.1g).

**Campbell CS300 Pyranometer**

The *CS300 Pyranometer* measures incoming and reflected shortwave radiation (Table 4.1) in the spectral length 300 nm to 1100 nm (CS, 2011a). It uses a silicon photovoltaic detector. The sensor is dome shaped to prevent water accumulation and constructed to eliminate internal condensation (Fig. 4.1g).

The sensor range is $0 \text{ W m}^{-2}$ to $2000 \text{ W m}^{-2}$ and has an accuracy of $\pm 5\%$ of the daily radiation. Every 15 min the mean and total incoming and reflected is recorded and the net is calculated (Table A.1).

The sensors was mounted on the south facing end of the crossarm (one is facing upwards and one downwards) on the station on Rabots glaciär.

**Campbell NR01 4-component net radiation sensor**

The the *NR01 4-component Net Radiation Sensor* consists of four sensors (Table 4.1). One up-facing and one downfacing *SR01 Pyranometer* measured incoming and reflected shortwave (305 nm to 2800 nm) radiation and one up-facing and one down-facing *IR01 Pyrgeometers* that measured incoming and outgoing longwave (4500 nm to 50 000 nm) radiation (CS, 2011b). The internal temperature of the NR01 is, when needed, automatically heated to reduce formation of dew and to melt frost (Fig. 4.1g).

The sensor measuring shortwave radiation has a range of $0 \text{ W m}^{-2}$ to $2000 \text{ W m}^{-2}$ and an accuracy of $\pm 10\%$ of the daily radiation. The mean and total incoming and reflected radiation was recorded and the net radiation and albedo was calculated every 15 min (Table A.1).

The sensor measuring longwave radiation has a range is $0 \text{ W m}^{-2}$ to $1000 \text{ W m}^{-2}$ and has an accuracy of $\pm 10\%$ of the daily radiation. Every 15 min the mean and total incoming and outgoing radiation is recorded and adjusted using an equation provided by the sensor manufacturer. The net longwave radiation is then calculated (Table A.1).

The sensors was mounted on the south facing end of the crossarm on the station on Storglaciären.

**Campbell CS CR1000 datalogger**

The *CS CR1000 datalogger* was controlled by a logger program (code listing A.1.2; A.1.2; A.1.2) that ran every 10 s and recorded every 15 min (Table A.1). During
periods of low incoming solar radiation the solar panel on Rabots glaciär did not manage to reload the battery and data was lost. On 14 July 2013 a new more energy economic logger program was installed.

### 4.1.2. Surface height

**Campbell SR50A, Sonic ranging sensor**

The *SR50A, Sonic Ranging Sensor* is an acoustic sensor that measured the distance to the ground (Table 4.1) by measuring the elapsed time between the emitted ultrasonic pulse and the return pulse (CS, 2011c). The raw distance was corrected for the air temperature according to the manufacturer (Table A.1).

The sensors range is 0.5 m to 10 m and has an accuracy of ±0.01 m or 0.4 % depending on which is greater.

The sonic ranger was coupled to the datalogger at the weather station but mounted separately on the crossarm of an aluminium stake drilled down into the ice. The position of the sensor is therefore constant in space making it possible to measure the ablation. Every 15 min the distance and the quality of the return signal was recorded. The quality of the return signal gave an estimation of the certainty of the value. At 0 the sensor was not able no make a measurement, at 152–210 the measurement is good, at 210–300 the signal is reduced and at 300–600 the measurement has a high uncertainty.

### 4.2. Data processing

#### 4.2.1. Field adjustments

A few measurements and notes had was taken in the field.

- The cardinal direction corresponding to the sensor 0° was measured. This measurement was used to convert the sensor output to cardinal directions.

- The stake where the sonic ranger was mounted needed to be redrilled to prevent the stake from melting out. The lowering of the sensor had to be compensated in the sensor output.

- When visiting the weather stations the surface properties at the site was noted. These were used to validate the albedo thresholds in the ablation calculations.

#### 4.2.2. Removing inaccuracies

Within the investigated period, individual or periods of diverging measurements was visible (Fig. A.2. After locating these measurements they were removed.

- The weather station on Storglaciären overturned due to hight wind velocities. All measurements within this period (19–23 Apr., 2013) was removed.

- All precipitation before the start of the melt season (16 May, 2013) was removed considering these measurement likely were melting snow accumulated in the catchment bucket. In the beginning of the melt season this could still be
the case but considering that air temperature was above melting point rainfall could not be ruled out.

- Both stations were installed in the end of winter. However, snow fell after the installation and the temperature and humidity probe at 0.5 m was buried in snow (Storglaciären, 14–17 May). On Rabots glaciär the sonic ranger was drilled several days after the station was installed. Consequently the amount of snow fallen after the stations installation is not known and the point of the burial had to be interpret from the behavior of the 0.5 m data compared to the 1 m data (Rabots glaciär, 24 Apr.–21 May).

- The quality signal recorded from the SR50 was used to remove noise from this sensor. The threshold was set to only using values between 152 and 300.

- The sonic ranger still showed several diverging measurements and were manually removed. The largest gap in the data from Storglaciären was from were the stake almost melted out of the ice and was therefore very unstable (from 30 July until redrilling, 8 August, 2013). The measurements recorded during the redrilling of the stake where the SR50 was mounted (22 July, 2013) on Rabots glaciär was also removed.

- Individual diverging measurements of temperature, relative humidity, radiation, wind speed and wind direction were removed manually.

### 4.2.3. Recalculate program errors

In the logger program for Storglaciären a few errors were discovered and had to be corrected.

- The totalized shortwave radiation was stored min\(^{-1}\) instead of s\(^{-1}\). The data was recalculated resulting in the unit J m\(^{-2}\).

- The albedo was calculated with the incoming shortwave radiation divided by the reflected radiation consequently giving albedo values over 1.0. The values was exchanged with values using the opposite division.

- The total radiation was incorrectly calculated. As mentioned above, to get the correct longwave radiation values, complementary calculations needs to be done from the original measurement. In the total net radiation the original measurement was used instead of the calculated value. The total net radiation was then recalculated using the correct values.

### 4.3. Data analyzing

#### 4.3.1. Basic statistics

To make the large data set more manageable and to lessen the influence of any remaining noise averages (mean temperature, relative humidity, shortwave radiation and wind speed) or totals (totalized radiation, precipitation and ablation) at different time intervals was used. To easier make comparisons between the glaciers and the
evolution of the meteorological parameters the data was broken into 3 h, daily and monthly means or totals. The 3 h interval was chosen to be able to get a manageable seasonal overview with diurnal fluctuations visible. Daily intervals were chosen as the shortest possible interval where the ablation should exceed the margin of error of the sonic ranger. Monthly intervals were chosen to visualize possible seasonal variation.

To be able to see the variation within the averages and to be able to compare the variation between the parameters the coefficient of variation was calculated for the monthly averages. The coefficient of variation is the standard deviation of the calculated mean divided by the mean and multiplied by 100.

To compare the fluctuation pattern of the meteorological parameter between the glaciers a correlation analysis was done. The correlation coefficient (R) is an index where 0 has no correlation and 1 has total correlation. The correlation can be negative or positive.

To investigate the meteorological parameters influence on ablation simple linear regression was done. The coefficient of determination ($r^2$) indicates how well the modelled line describes the data where 0 is not at all, and 1 is linear. When calculating how ablation respond to the different meteorological parameters the $y$-intersect can be used to see the ablation when the parameter is 0. When $r^2$ is significant the slope of the line can be used as an ablation coefficient. A simple linear regression was also done when investigating the relationship between wind speed and the wind speed standard deviation.

### 4.3.2. Calculations

#### Melt season

Considering that ablation is one of the main focuses for the study, several calculations only include the melt season. The melt season is set from where both glaciers had a stable daily mean temperature above melting point and melt had become evident. This began the 16th of May and ended when the first weather station was removed (5 September, 2013).

#### Temperature and relative humidity

To get a monthly value of temperature that only describes periods with ablation the positive degree days were calculated. For every month the daily mean temperatures above 0 °C was summarized. To neutralize the different amount of days in a month, the sums was divided by the amount of days for that month and multiplied with 30.

To get a better understanding of the change of vapor in the air the vapor pressure was calculated from the temperature and relative humidity.

For temperature, relative humidity and vapor pressure vertical gradients were calculated. The value for 0.5 m above the surface was subtracted from the value for 2 m and then divided by 1.5 resulting in the units °C m$^{-1}$, % m$^{-1}$ and Pa m$^{-1}$, respectively. Thereafter daily and monthly averages was calculated. To visualize the change from the lower to the middle sensor and finally to the higher sensor calculation of the difference between the sensors values was calculated and plotted against the surface height.
Throughout the report the values for the 0.5 m and the 2 m sensors were used and the 1 m sensors often neglected. A focus on the 0.5 m sensors was chosen due to its proximity to the processes at the atmosphere-glacier interface and the 2 m sensors was chosen because it is likely to be the least affected by the atmosphere-glacier interface. The highest sensor is also interesting because its the height were the wind, radiation and precipitation is measured.

Radiation

For the sensors measuring shortwave and longwave radiation on Storglaciären the total radiation was not stored and needed to be calculated. The 15 min mean was multiplied by 900 (amount of seconds in 15 min) to convert the measured value in W m\(^{-2}\) to J m\(^{-2}\).

The most important information obtained by shortwave radiation measurements for a melting glacier is the amount of energy it brings to the system. However, to investigate reasons behind lower or higher measurements, incoming shortwave radiation can be used to study general cloud conditions. For a study stretching over a longer period of time the shape of the daily radiation curve tell more than the value itself considering an overcast day in summer is likely to contribute more energy than a clear day in winter. To estimate cloudiness the shape of the daily incoming radiation curve were studied and categorized (Fig. 4.2). The curve has a very specific shape under a clear sky but get more difficult to separate with increasing cloud fraction and thickness. However it will give an approximation of the cloudiness on the glaciers when acknowledged methods using incoming longwave radiation (Giesen et al., 2009) is not possible.

\[0\quad 200\quad 400\quad 600\quad 800\]
\[\text{Clear}
\text{Partly cloudy}
\text{Cloudy}
\text{Thick cloud layer}\]
\[I \varnothing \text{(Wm}^{-2}\text{)}\]

*Figure 4.2.* Characterization of the amount of clouds from the curve of the incoming shortwave radiation.

Precipitation

Precipitation was mainly studied using calculated 3 h, daily and monthly totals. In some analysis however, it was more reasonable to separate measurements of a significant amount of rain from the numerous measurements of extremely low amounts. This was done by setting a threshold that excluded low values. The threshold was chosen using a definition for when drizzle becomes rain (0.5 mm h\(^{-1}\)) defined by Swedish Meteorological and Hydrological Institute (SMHI).
Wind speed

Wind speed was studied using daily, monthly and seasonal averages. The values are mainly presented as wind roses (section 5.1.4). The measured wind directions are divided into a specific amount \((n)\) of main directions (for this study, \(n = 8\) i.e., north, north-east, east, south-east, south, south-west, west, north-west). The fraction of the amount of measurements within every direction decides the length and consequently width of a sector. The \(n\) sectors make up a circle that illustrate the cardinal directions. In a wind rose it is possible to include a second parameter that describes a property of the air (e.g. velocity, temperature or relative humidity). Within every sector the values for the second parameter is divided into color coded intervals and are plotted, like a stacked bar diagram, within the sector. This results in a figure that illustrates the distribution of the wind direction and the property of the air. This can be used to investigate if one air property is more common to arrive with air from a specific direction.

The measured standard deviation of the mean 15 min wind speed data can be used as an indication of the stability of the wind speed. Low standard deviation indicate a constant value of mean wind speed whereas higher standard deviation values indicates a gusty wind.

Turbulent Heat Fluxes

The transfer of latent and sensible heat make up the turbulent heat flux. Convection and conduction are the major processes involved and consequently the temperature and humidity are the meteorological parameters that influence the turbulent heat fluxes most and should be enhanced by air turbulence (Wheler, 2009).

The bulk aero-dynamic method (e.g. Oerlemans, 2001; Hock, 2005; Wheler, 2009) is based on the fact that a melting surface always has the temperature of 0 °C and a constant vapor pressure of 611 Pa (Munro, 1990). The method uses a transfer coefficient and a stability correction that needs further knowledge about the glacier atmosphere interactions close to the surface. An approach based on the bulk aerodynamic method but applicable for scenarios when a glacier wind is dominant and the turbulence is not well known, is developed for valley glaciers by Oerlemans (2010). The sensible \((H_s)\) and the latent \((H_l)\) heat flux is calculated as follows:

\[
H_s = \rho c_p C^* (T_z - T_s) \quad (4.1)
\]
\[
H_l = \rho L_v C^* (q_z - q_s) \quad (4.2)
\]

where \(\rho\) is the air density, \(c_p\) is the specific heat, \(C^*\) is a turbulent exchange coefficient, \(T_s\) is the temperature at the surface, \(T_z\) the temperature at height \(z\), \(L_v\) is the latent heat of evaporation, \(q_s\) is the specific humidity at the surface and \(q_z\) is the specific humidity height \(z\). The specific humidity \((q_z - q_s)\) is calculated from the vapor pressure and relative humidity using:

\[
(q_z - q_s) = \frac{0.622}{P} (e_z - e_s) \quad (4.3)
\]

where \(P\) is the atmospheric pressure, \(e_s\) is the vapor pressure at the surface and \(e_z\) the vapor pressure at height \(z\).

\[
e_z = \frac{R H e_0}{100} \quad (4.4)
\]
where RH is the relative humidity and $e_0$ the saturation water pressure of the air:

$$e_0 = 611.213 \exp(17.5043 \frac{T_z}{T_z + 241.2})$$

were $T_z$ is the measured temperature. The turbulent exchange coefficient, $C^*$, is differing depending on the temperature gradient:

$$C^* = C_b + C_{kat} = C_b + k(T_z - T_s)\sqrt{\frac{g}{T_0\gamma\theta Pr}}$$

for $T_z - T_s > 0$ (4.6)

and

$$C^* = C_b$$

for $T_z - T_s \leq 0$ (4.7)

were $T_s$ is the surface temperature, $C_b$ is contribution associated with the turbulence generated by the synoptic atmospheric circulation and $C_{kat}$ is contribution from the katabatic wind system and carry the values:

$$C_b = 0.003, k\sqrt{\frac{g}{T_0\gamma\theta Pr}} = 0.0002$$

(4.8)

**Water equivalent ablation**

To calculate the amount of ablation in water equivalent, knowledge of the surface properties and an approximation of water equivalent for snow, slush or water was needed. To do this the albedo was analyzed. As mentioned above, the albedo of snow and ice differs significantly and can be used as an indication of the surface properties of the glacier. The albedo was calculated using:

$$\alpha = \frac{SW_{out}}{SW_{in}}$$

Jonsell et al. (2003) did an extensive study on the albedo on Storglaciären where only values when the zenith angle was $< 65^\circ$ to avoid periods with low shortwave radiative fluxes. The behavior of the albedo can be seen in appendix A.2. Since the albedo is relatively stable in the daytime but diverge significantly during nights daily averages between the hours 10.00 and 14.00 was calculated and used as daily values for albedo. Figure 4.3a and c show the daily albedo values over the investigated period and figure 4.3b and d show the incoming radiation against the reflected shortwave radiation. In the latter plot, two almost linear patterns can be seen. These are interpret as the reflection of snow and ice, respectively (Oerlemans and Grisogono, 2002) and the thresholds are chosen considering. The space between is interpret as slush and the points above the black 1:1 line was likely caused by snowfall that covered the upfacing sensor (Oerlemans, 2010). The water equivalent for snow ($W_S$) was estimated to 90%, 55% for ice ($W_I$) and 70% for slush ($W_{Sl}$). The water equivalent ablation is then calculated:

$$M_{weq} = \text{for } \alpha > \alpha_s \Rightarrow W_SM, \quad \alpha < \alpha_i \Rightarrow W_IM, \quad \alpha_i > \alpha < \alpha_s \Rightarrow W_{Sl}M$$

(4.10)

where $\alpha$ is the measured albedo, $\alpha_s$ the albedo threshold for snow, $\alpha_i$ the albedo threshold for ice and $M$ the measured ablation in m.
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation

Figure 4.3. a) evolution of the albedo on Rabots glaciär b) incoming shortwave radiation against reflected shortwave radiation at Rabots glaciär c) evolution of the albedo on Storglaciären d) incoming shortwave radiation against reflected shortwave radiation at Storglaciären.

4.3.3. List of variable names

In figures and tables variable names are used. All calculations were made for specific time intervals which are presented every time a variable name is used. $z$ specifies the height above the surface (m) where the sensor was mounted. This is only specified if multiple sensors are installed at different heights.
$T_z$ Average temperature (°C)
$PDD_z$ Positive degree days (°C)
$max \ T_z$ Highest temperature recorded (°C)
$min \ T_z$ Lowest temperature recorded (°C)
$T_z - T_{0.5}$ Average temperature gradient (°C m$^{-1}$)
$Rh_z$ Average relative humidity (%)
$min \ Rh_z$ Lowest relative humidity recorded (%)
$Rh_z - Rh_{0.5}$ Average relative humidity gradient (% m$^{-1}$)
$\bar{e}_z$ Average vapor pressure (Pa)
$\bar{H}_{(st)}$ Average heat fluxes ($s$ specifies sensible heat and $l$ latent heat) (W m$^{-2}$)
$\sum H_{(st)}$ Total turbulent heat fluxes ($s$ specifies sensible heat and $l$ latent heat) (J m$^{-2}$)
$I_\downarrow$ Incoming shortwave radiation (W m$^{-2}$)
$I_\uparrow$ Reflected shortwave radiation (W m$^{-2}$)
$I_\downarrow - I_\uparrow$ Average net shortwave radiation (W m$^{-2}$)
$\sum (I_\downarrow - I_\uparrow)$ Total net shortwave radiation (J m$^{-2}$)
$\alpha_n$ Average albedo at noon (10:00–14:00)
$\bar{\alpha}$ Albedo
$s_w$ Average wind speed (m s$^{-1}$)
$max \ s_w$ Highest wind speed recorded (m s$^{-1}$)
$avg \ max \ s_w$ Average highest wind speed recorded (m s$^{-1}$)
$\sigma s_w$ Average standard deviation of the wind speed recorded (m s$^{-1}$)
$max \ \sigma s_w$ Average highest standard deviation of the wind speed recorded (m s$^{-1}$)
$max \ \sigma s_w$ Highest standard deviation of the wind speed recorded (m s$^{-1}$)
$\sum P$ Total amount of rain fall (mm)
$\Delta h$ Change in surface height (mm)
$\sum \Delta h_{m.w.e.}$ Total daily water equivalent change in surface height (mm)
5. Results

5.1. Meteorological overview

5.1.1. Temperature

The temperature was measured with the Campbell HC2S3, Temperature and relative humidity probe, at 0.5 m, 1 m and 2 m above the glacier surface. The values presented beneath are 3-hour, monthly and seasonal averages from the 0.5 m and 2 m sensors. The 1 m sensor was excluded to make the 0.5 m and 2 m more distinct. Figure 5.1a–b show the relationship between the 1 m and the 2 m sensor. The y-intersect on both glaciers was −0.2 °C and the value for the 1 m sensor was generally 99 % and 95 %, respectively of the value of the 2 m sensor.

Figure 5.1. Plot illustrating the relationship between values of the a) 2 m and 1 m temperature on Rabots glaciär, b) 2 m and 1 m temperature on Storglaciären, c) 2 m and 1 m relative humidity on Rabots glaciär, d) 2 m and 1 m relative humidity on Storglaciären.

Rabots glaciär

Figure 5.2 illustrates the mean temperature at Rabots glaciär for the studied period. Until the middle of April the temperature was below zero. During this time the temperature at 0.5 m fluctuated around the 2 m temperature. From the middle
of April to the middle of May the lower sensor was buried in the snowpack. Consequently the 2 m sensor was only 1.5 m above the surface at this time. From the beginning of May the temperature was below zero but increased steadily until the middle of May when the temperature rose above zero. Throughout the season there were periods where temperature increased relative quickly, peaked for a week or two and quickly fell, sometimes below freezing point. The difference between the 2 m and 0.5 m sensors was larger during these peaks.

**Figure 5.2.** Plot showing the 3-hour mean temperature at 0.5 m and 2 m above the surface at Rabots Glaciär, 2013

Table 5.1 show mean values for the different meteorological parameters for each month and the melt season. No mean values are calculated for the 0.5 m sensor in May because loss of data prevents formation of accurate means. The mean temperature at 0.5 m and 2 m in June was 3.6°C and 4.8°C, respectively and stayed relatively constant in July (3.5°C and 4.8°C, respectively) and lowered slightly in August with 3.2°C at 0.5 m 4.4°C at 2 m. As seen in table 5.1 the coefficient of variation at both heights was higher in the beginning of the season and decreased every month. Generally the coefficient of variation was lower at the 2 m sensor. The difference between the positive degree days at 0.5 m and 2 m for the individual months differs little from the mean temperature pattern. The difference in positive degree days between the sensors in June was 35°C, 39.5°C in July and 38.3°C in August. May was the month with both the highest and the lowest values recorded on the glacier.

The temperature gradient from 0.5 m to 2 m above the surface was relatively even for the months (Table 5.1). In June the gradient was 0.8°C m⁻¹ in July it was 0.9°C m⁻¹ and in August 0.8°C m⁻¹. However, the coefficient of variation was relatively large for all months (75.6% to 101.2%). The mean gradient for the melt season was 0.9°C m⁻¹.
Table 5.1. Averages, totals, highest and lowest values and coefficient of variation (cv) for meteorological parameters measured from 1 May to 31 August, 2013 at Rabots Glaciär (gray rows) and Storglaciären (white rows). Values are rounded to the nearest integer. For variable names see section 4.3.3.

<table>
<thead>
<tr>
<th>Variable</th>
<th>MAY cv</th>
<th>JUN cv</th>
<th>JUL cv</th>
<th>AUG cv</th>
<th>SEASON[^2]</th>
</tr>
</thead>
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<tr>
<td>$T_{0.5m}$</td>
<td>− − 79.2</td>
<td>− − 3.6</td>
<td>− − 3.8</td>
<td>− − 4.1</td>
<td>− − 4.1</td>
</tr>
<tr>
<td>$T_{2m}$</td>
<td>1.6 72.4</td>
<td>1.6 54.7</td>
<td>1.6 4.1</td>
<td>1.6 4.1</td>
<td>1.6 4.1</td>
</tr>
<tr>
<td>PDD$_{0.5m}$</td>
<td>− − 110.7</td>
<td>− − 5.2</td>
<td>− − 113.5</td>
<td>− − 5.6</td>
<td>− − 5.6</td>
</tr>
<tr>
<td>PDD$_{2m}$</td>
<td>95.6 147.5</td>
<td>95.6 137.5</td>
<td>95.6 126.8</td>
<td>95.6 447.3</td>
<td>95.6 541.2</td>
</tr>
<tr>
<td>max $T_{2m}$</td>
<td>15.2 14.6</td>
<td>16.7 14.6</td>
<td>− − 14.6</td>
<td>− − 14.6</td>
<td>− − 15.2</td>
</tr>
<tr>
<td>min $T_{2m}$</td>
<td>− − 3.9</td>
<td>− − −1.1</td>
<td>− − −0.6</td>
<td>− − −0.6</td>
<td>− − −1.3</td>
</tr>
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<td>$T_{2m} - T_{0.5m}$</td>
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<td>− − 1.0</td>
<td>− − 1.0</td>
<td>− − 1.0</td>
<td>− − 1.0</td>
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<tr>
<td>$R h_{0.5m}$</td>
<td>− − 13.4</td>
<td>− − 12.8</td>
<td>− − 12.8</td>
<td>− − 12.8</td>
<td>− − 12.8</td>
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<tr>
<td>$R h_{2m}$</td>
<td>73.4 55.5</td>
<td>74.7 55.9</td>
<td>75.7 55.9</td>
<td>76.7 55.9</td>
<td>75.7 75.9</td>
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<tr>
<td>min $R h_{0.5m}$</td>
<td>33.3 55.3</td>
<td>33.3 55.3</td>
<td>33.3 55.3</td>
<td>33.3 55.3</td>
<td>33.3 55.3</td>
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<tr>
<td>$R h_{2m} - R h_{0.5m}$</td>
<td>− − 2.4</td>
<td>− − −2.2</td>
<td>− − −2.2</td>
<td>− − −2.2</td>
<td>− − −2.2</td>
</tr>
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<td>$I_\downarrow$</td>
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<td>129.3 150.8</td>
<td>129.3 150.8</td>
<td>129.3 150.8</td>
<td>129.3 150.8</td>
</tr>
<tr>
<td>$I_\uparrow$</td>
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<td>121.3 154.2</td>
<td>121.3 154.2</td>
<td>121.3 154.2</td>
<td>121.3 154.2</td>
</tr>
<tr>
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<td>154.3 321.6</td>
<td>154.3 227.5</td>
<td>154.3 227.5</td>
<td>154.3 227.5</td>
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<td>2.7 2.8</td>
<td>2.7 2.8</td>
<td>2.7 2.8</td>
</tr>
<tr>
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<td>19.4 23.5</td>
<td>19.4 23.5</td>
<td>19.4 23.5</td>
<td>19.4 23.5</td>
</tr>
<tr>
<td>$s_w \sigma$</td>
<td>0.8 28.7</td>
<td>0.8 28.7</td>
<td>0.8 28.7</td>
<td>0.8 28.7</td>
<td>0.8 28.7</td>
</tr>
<tr>
<td>max $s_w \sigma$</td>
<td>4.8 4.6</td>
<td>4.8 4.6</td>
<td>4.8 4.6</td>
<td>4.8 4.6</td>
<td>4.8 4.6</td>
</tr>
<tr>
<td>$\sum P[^3]$</td>
<td>6.0 526.0</td>
<td>6.0 526.0</td>
<td>6.0 526.0</td>
<td>6.0 526.0</td>
<td>6.0 526.0</td>
</tr>
<tr>
<td>$\Delta h$</td>
<td>34.0 782.2</td>
<td>34.0 782.2</td>
<td>34.0 782.2</td>
<td>34.0 782.2</td>
<td>34.0 782.2</td>
</tr>
<tr>
<td>$\sum \Delta h$</td>
<td>703.3 3838.5</td>
<td>703.3 3838.5</td>
<td>703.3 3838.5</td>
<td>703.3 3838.5</td>
<td>703.3 3838.5</td>
</tr>
</tbody>
</table>


Storglaciären

Figure 5.3 illustrates the mean temperature at Storglaciären for the studied period. The weather station was not installed until the middle of April. Shortly after the installation the station blew over and was down for a few days. These measurements were removed. The temperature at 0.5 m was constantly lower than the temperature...
at 2 m. During a short period in the middle of May the lower sensor was buried in the snow pack. Consequently the 2 m sensor was at this time only 1.5 m above the surface.

From the middle of May throughout the season the temperature mainly stayed above zero. Constantly during the season there were periods where temperature increases relative quickly, peaked for a week or two and quickly fell, sometimes to sub-zero temperatures. When the temperature peaked the difference between the 2 m temperature and the 0.5 m temperature was greater. These peaks were more distinct in the beginning and end of the melt season.

![Figure 5.3](image)

**Figure 5.3.** Plot showing the 3-hour mean temperature at 0.5 m and 2 m above the surface at Storglaciären, 2013. The axes agree with the corresponding figure for Rabots glaciär but the station on Storglaciären was installed later in the season.

As seen in table 5.1 mean temperature for the melt season was 4.1 °C at 0.5 m and 5.6 °C at 2 m. The mean temperature for the individual months are relatively even. In June the mean temperature was 3.8 °C and 5.2 °C, respectively, in July 4.1 °C and 5.6 °C, respectively and in August 4.1 °C and 5.6 °C, respectively. Generally the coefficient of variation was lower at 2 m and the value decreased every month. The difference between the positive degree days at 0.5 m and 2 m for the individual months was similar to the mean temperature pattern. The difference in positive degree days between the sensors for June was 43.7 °C, 47.6 °C in July and 45.5 °C in August. May was the month with both the highest and the lowest measurement recorded on the glacier.

The mean temperature gradient from 0.5 m to 2 m was 1.0 °C m⁻¹ for all investigated months (Table 5.1). The coefficient of variation was relatively large for all months (69.6% to 84.3%).

### 5.1.2. Relative humidity

The relative humidity was measured with the Campbell HC2S3, Temperature and relative humidity probe, at 0.5 m, 1 m and 2 m above the glacier surface. The values
presented beneath are 3-hour, monthly and seasonal averages from the 0.5 m and 2 m sensors. The 1 m sensor was excluded to make the 0.5 m and 2 m more distinct. Figure 5.1c–d show the relationship between the 1 m and the 2 m sensor. The y-intersect on Rabots glaciär and Storglaciären was 6.3 % and 7.2 %, respectively and the value for the 1 m sensor was generally 95 % and 88 %, respectively of the value of the 2 m sensor.

**Rabots glaciär**

Figure 5.4 illustrates the mean relative humidity at Rabots glaciär for the studied period. In the first half of May the relative humidity measured at 0.5 m fluctuated between 80 % to 10 % whereas the relative humidity at 2 m fluctuated from below 40 % to 100 %. From the middle of May to the middle of June the lower sensor was under the snow. Throughout the season the relative humidity fluctuated and in late May and late July there was longer period of low relative humidity. The difference between the sensors are small but increased when the humidity dropped. Generally the relative humidity at 0.5 m was higher than at 2 m.

Table 5.1 show the mean relative humidity for the individual months and the melt season. The mean relative humidity for the melt season was 85.9 % at 0.5 m and 81.5 % at 2 m. The difference between the individual months was small (84.9 % to 89.1 % at 0.5 m and 81.2 % to 85.9 % at 2 m). The coefficient of variation was relatively low at both sensors.

The relative humidity gradient did not change significantly during the season. A decrease of relative humidity with height can be seen throughout the season with values from $-2.2 \% m^{-1}$ to $-2.4 \% m^{-1}$. The mean relative humidity gradient for the melt season was $-2.6 \% m^{-1}$.

![Figure 5.4](image-url)

*Figure 5.4.* Plot showing the 3-hour mean relative humidity at 0.5 m and 2 m above the surface at Rabots Glaciär, 2013.
Storglaciären

Figure 5.5 illustrates the mean relative humidity at Storglaciären for the studied period. The weather station was not installed until the middle of April. Shortly after the installation the station blew over and was down for a few days. These measurements were removed. During a short period in the middle of May the lower sensor was buried in the snow pack. Throughout the season the relative humidity fluctuated from 60% to 100%. The relative humidity at 0.5 m was generally slightly higher than at 2 m. In late May and late July there was a period of lower relative humidity. At these periods the difference between the sensors was greater.

The mean relative humidity for the individual months were similar. The relative humidity in June was 85.0% at 0.5 m and 81.5% at 2 m, in July it was 83.7% and 79.9%, respectively and in August it was 83.7% and 80.5%, respectively. The mean relative humidity for the melt season was 82.7% and 78.5%, respectively. The coefficient of variation was relatively low at both sensors.

The relative humidity gradient does not change significantly during the season. A decrease with height can be seen throughout the season with values from $-2.2\%\,\text{m}^{-1}$ to $-2.5\%\,\text{m}^{-1}$. The mean relative humidity gradient for the melt season was $-2.5\%\,\text{m}^{-1}$.

![Figure 5.5](image-url)

*Figure 5.5.* Plot showing the 3-hour mean relative humidity at 0.5 m and 2 m above the surface at Storglaciären, 2013. The axes agree with the corresponding figure for Rabots glaciar but the station on Storglaciären was installed later in the season.

5.1.3. Radiation

Rabots glaciar

Figure 5.6 show the 3-hour mean incoming shortwave radiation measured with a Campbell CS300 Pyranometer at 2 m above the glacier surface. The incoming shortwave radiation had a daily fluctuation pattern. From the middle of May to late June the daily radiation fluctuation does not reach down to zero before increasing again.
Missing data can be seen in early May when low battery and insufficient solar radiation caused logger failure and the logger was not able to record data during night. There were periods of relatively high incoming shortwave radiation in late May and late July with values almost up to 700 W m$^{-2}$. Generally the shortwave radiation from June throughout the season was low.

The mean incoming radiation for the melt season was 148.7 W m$^{-2}$ and August was the month with the lowest mean value (98.2 W m$^{-2}$) (Table 5.1). Some days the incoming shortwave radiation did not exceed 100 W m$^{-2}$. In May the total net shortwave radiation was 99.4 MJ m$^{-2}$, in June 129.4 MJ m$^{-2}$, in July 182.6 MJ m$^{-2}$ and in August 142.4 MJ m$^{-2}$. The coefficient of variation is relatively high throughout the season.

The albedo changed significantly over the studied period. In May the albedo was 0.8, in June 0.7, in July 0.5 and in August 0.4. Figure 4.3 show the evolution of the surface albedo. In late June the albedo dropped significantly and for the rest of the season it was relatively stable on a lower level. A few high peaks can be seen in July and late August.

![Figure 5.6. Plot showing the 3-hour mean incoming shortwave radiation on Rabots Glaciär, 2013.](image-url)

**Storglaciären**

Figure 5.7 show the 3-hour mean incoming solar radiation measured with a SR01 Pyranometer at 2 m above the glacier surface. The incoming shortwave radiation had a daily fluctuation pattern. From the middle of May to late June the daily radiation fluctuation does not reach down to zero before increasing again. In the middle of May and late July there were periods of relatively high incoming shortwave radiation with values over 700 W m$^{-2}$. Generally the shortwave radiation from June throughout the season was low.

The mean incoming radiation for the melt season was 181.2 W m$^{-2}$ and August was the month with the lowest mean value (150.8 W m$^{-2}$) (Table 5.1).
Figure 5.7. Plot showing the 3-hour mean incoming shortwave radiation on Storglaciären, 2013. The axes agree with the corresponding figure for Rabots glaciär but the station on Storglaciären was installed later in the season.

Some days the incoming shortwave radiation did not exceed 100 W m$^{-2}$. In May the total incoming solar radiation was 154.3 MJ m$^{-2}$, in June 215.0 MJ m$^{-2}$, in July 321.6 MJ m$^{-2}$ and in August 227.5 MJ m$^{-2}$. The coefficient of variation was relatively high throughout the season.

The albedo changed significantly over the studied period. In May the albedo was 0.8, in June 0.5, in July 0.3 and in August 0.4. In figure fig:albedo the evolution of the surface albedo can be seen. In late June the albedo dropped significantly and stabilized on a lower level. A few peaks in albedo can be seen in July and late-August.

5.1.4. Wind direction and speed

The direction and speed of the surface boundary wind was measured with a Campbell 05103 Wind monitor at 2 m above the surface at Rabots glaciär and Storglaciären. The wind direction and speed are presented in the form of wind roses (subsection 4.3.2). Figure 5.8 show the distribution of wind directions and wind speed on Rabots glaciär and Storglaciären.

Rabots glaciär

In the measured period, 23 % of the winds at Rabots glaciär were north-easterly and 24 % easterly. These winds were seldom higher than 4 m s$^{-1}$. South-westerly winds make up 20 % of the directions and contained the highest wind speeds recorded on the glacier.

Figure 5.9 show the monthly differences in wind direction and air properties, such as velocity, temperature and relative humidity, on Rabots glaciär from 1 April to 31 August, 2013. In April, 26 % of the wind were north-easterly and 23 % arrived from the opposite direction. The south-westerly winds had the highest velocity.
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation

No significant difference in temperature can be seen but the south-westerly winds were more humid. In May the main winds were north-easterly (25 %) and easterly (20 %). The stronger winds were south-westerly (15 %) and these winds lacked air temperatures above 10°C and relative humidity under 80 %. A few percent of the winds that reached 9\,ms\textsuperscript{-1} were northerly winds. In June, an even larger amount of winds were easterly (30 %). North-easterly winds made up 24 % and south-westerly winds 13 %. The south-easterly winds was more humid and lacked higher temperature but no higher wind speeds were recorded. The only winds that reached wind speeds up to 8\,ms\textsuperscript{-1} were a few percent northerly winds. In July wind speeds up to 15\,ms\textsuperscript{-1} of south-westerly winds were recorded. This air lacked higher temperatures and lower humidity and made up 24 % of the winds. 29 % of the winds were easterly and 16 % north-easterly. In August the south-westerly winds made up 23 % and were the highest recorded wind speeds. No great differences in air temperature can be seen but the south-easterly winds lacked low humidity. 27 % of the winds recorded were easterly and 24 % north-easterly.

Storglaciären

On Storglaciären approximately 46 % of the winds were westerly, 13 % south-westerly. All winds with velocity higher than 6\,ms\textsuperscript{-1} were westerly and south-westerly. 11 % of the winds were easterly and these never exceeded 6\,ms\textsuperscript{-1}.

Figure 5.10 show the monthly differences in wind direction and air properties on Storglaciären for the period 18 April to 31 August, 2013. In April, 48 % of the winds were westerly and a smaller amount easterly (12 %), south-easterly (14 %) and south-westerly (13 %). The westerly winds were the only winds with velocity higher than 6\,ms\textsuperscript{-1} and together with south-westerly winds they were the only winds with relative humidity under 80 %. May was the least homogenous month. The distribution of wind directions were, westerly winds (34 %), south-westerly (15 %), easterly (14 %) and south-easterly (10 %). The recorded western wind contained the highest wind speed, the dryest and warmest air. In June 48 % were westerly and remaining
Figure 5.9. Wind roses for Rabots glaciär, 1 Apr.–5 Sep., 2013. Wind roses that divide the measured wind directions into 8 directions. The fraction of the amount of measurements within every direction decides the length and consequently width of a sector. The 8 sectors make up the circle that illustrate the cardinal directions (north, north-east, east, south-east, south, south-west, west, north-west). Wind speed, temperature and relative humidity is plotted as a second parameter that is divided into color coded intervals and plotted as a stacked bar diagram within the sector.
Figure 5.10. Wind roses for Rabots glaciär, 18 Apr.–5 Sep., 2013. Wind roses that divide the measured wind directions into 8 directions. The fraction of the amount of measurements within every direction decides the length and consequently width of a sector. The 8 sectors make up the circle that illustrate the cardinal directions (north, north-east, east, south-east, south, south-west, west, north-west). Wind speed, temperature and relative humidity is plotted as a second parameter that is divided into color coded intervals and plotted as a stacked bar diagram within the sector.
directions never exceed 10 % and lacked warm and dry air. In July 53 % of the winds were westerly and only these, together with the south-westerly winds, contain wind speeds higher than $6 \text{ m s}^{-1}$. Again this air was the driest and least cold. In August 53 % of the winds were westerly and 18 % south-westerly. At this time the westerly winds were the only ones that occasionally had air temperature above 10°C.

5.1.5. Precipitation

The precipitation was measured with a Campbell 52202, Tipping bucket raingauge, at 2 m above the glacier surface. The values presented beneath are 3-hour sums.

Rabots glaciär

The precipitation fallen on Rabots glaciär for the studied period can be seen in figure 5.11. The precipitation usually fell in shorter periods distributed over the entire melt season. The total precipitation for the melt season at Rabots glaciär was 526.0 mm (Table 5.1). In May 6.0 mm fell over 10 h, in June 168.3 mm over 98 h, in July 220.4 mm over 137 h and August 118.0 mm over 81 h (Table 5.1). Figure 5.12 show from which direction precipitation over 0.1 mm ($15 \text{ min}$) came. On Rabots glaciär the origin of the precipitation was spread in several directions. 18 % from south-west, 20 % from east, 17 % from north-east, 14 % from north. SMHI defines a normal rainshower as 1 mm to 10 mm of rain in 10 min to 20 min. The total amount of precipitation fallen during rain and rainshowers was 466.3 mm. Most of the precipitation is under 1 mm in 15 min. The largest amount of showers came from north.
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation

Figure 5.12. Wind roses for a) Rabots glaciär and b) Storglaciären. Wind roses that divide the measured wind directions into 8 directions. The fraction of the amount of measurements within every direction decides the length and consequently width of a sector. The 8 sectors make up the circle that illustrate the cardinal directions (north, north-east, east, south-east, south, south-west, west, north-west). Precipitation divided into color coded intervals and plotted as a stacked bar diagram within the sector.

Storglaciären

The precipitation fallen on Storglaciären for the studied period can be seen in figure 5.13. The precipitation usually fell in shorter periods distributed over the entire melt season. However in the middle of July a period stands out with values up to 8.7 mm (15 min)$^{-1}$. The total precipitation for the melt season at Storglaciären was 782.2 mm (Table 5.1). In May 34.0 mm rain fell over 53 h, in June 164.8 mm over 104 h, in July 365.0 mm over 144 h and in August 208.2 mm over 118 h (Table 5.1). Figure 5.12 show that on Storglaciären the origin of the precipitation over

Figure 5.13. Plot showing the 3-hour total precipitation at 0.5 m and 2 m above the surface at Storglaciären, 2013. The numbers in July are values that exceed the limit of the y-axis.

0.1 mm/15 min was spread in three main directions. 34% from west, 21% from east, 20% from south-west. The total amount of precipitation fallen during rain and rainshowers was 726.1 mm. Most of the precipitation was under 1 mm in 15 min
but 7% of the measurements were showers from west and 6% were showers from east.

5.2. Meteorological differences

5.2.1. Seasonal overview

Figure 5.14 show the temperature for both glaciers throughout the investigated period. On both glaciers the temperature measured at 2 m above the surface was higher than the temperature at 0.5 m and the fluctuation pattern were synchronized between the glaciers (R = 0.86 at 0.5 m and R = 0.96 at 2 m) (Fig. 5.15a–b). The mean temperature for the melt season at 2 m was almost one degree higher on Storglaciären (Rabots glaciär 4.8 °C and Storglaciären 5.6 °C) but only half a degree at 0.5 m (Rabots glaciär 3.6 °C and Storglaciären 4.1 °C). Consequently, the vertical temperature gradient was slightly higher on Storglaciären (melt season average of 0.9 °C m⁻¹ on Rabots glaciär and 1.0 °C m⁻¹) on Storglaciären. The positive degree day approach give similar results. On Storglaciären the positive degree days was 13 % higher at 0.5 m and 16 % higher at 2 m.

During the melt season the mean relative humidity was higher on Rabots glaciär (Fig. 5.14). At 0.5 m Rabots glaciär had 85.9 % and Storglaciären had 82.7 % and at 2 m Rabots glaciär had 81.5 % and Storglaciären had 78.5 % (Table 5.1). There was a correlation for the relative humidity fluctuations between the glaciers (R = 0.64 at 0.5 m and R = 0.67 at 2 m) (Fig. 5.15c–d).

The correlation in the fluctuation of vapor pressure was very good (R = 0.90 at 0.5 m and R = 0.95 at 2 m) between the glaciers (Fig. 5.15e–f).

The total net radiation (Fig. 5.14) is significantly higher on Storglaciären (Rabots glaciär 541.7 MJ m⁻² and Storglaciären 905.4 MJ m⁻² (Table 5.1). There is a correlation of the net shortwave radiation fluctuations (R = 0.79) and the incoming shortwave radiation fluctuation (R = 0.67) between the glaciers (Fig. 5.15h–i).

The albedo differs significantly between the glaciers. The mean albedo for the melt season at Rabots glaciär was 0.6 and on Storglaciären 0.4. The threshold for the snow albedo was set to 0.71 compared to 0.59 (Fig. 4.3) and the ice albedo was set to 0.51 compared to 0.39. The correlation of albedo between the glaciers make three clear clusters in a line (Fig. 5.15g) with a relatively good correlation (R = 0.89).

The mean wind speed and the mean maximum wind gust was higher on Storglaciären (Rabots glaciär 2.5 m s⁻¹ and Storglaciären 3.0 m s⁻¹, Rabots glaciär 4.4 m s⁻¹ and Storglaciären 4.9 m s⁻¹) (Table 5.1). The mean measured standard deviation was higher at Storglaciären and so was the seasons highest measured values (Rabots glaciär 23.5 m s⁻¹ and 28.7 m s⁻¹). Peaks in wind speed seems to occur at roughly the same time (Fig. 5.16) but the correlation was relatively low (R = 0.47 ) (Fig. 5.15k). During the peaks, the values for Storglaciären were usually higher and stayed high for a longer period of time. The standard deviation for the wind speed was higher on Storglaciären. During periods of lower wind speeds Rabots glaciär had small peaks whereas Storglaciären was relatively stable (Fig. 5.16). The periods with high wind speed coincide with periods where the wind direction on the glaciers were similar (Fig. 5.16).

The total precipitation for the melt season was 526.0 mm (338 h of rain) on Rabots
Figure 5.14. Illustration of daily values of meteorological parameters. a) mean temperature, b) mean relative humidity c) shortwave radiation, d) total precipitation at Rabots glaciär (blue) and Storglaciären (orange) from the beginning of April to early September, 2013.

glaciär and 782.2 mm (397 hour of rain) on Storglaciären (Table 5.1). The precipitation often occurred at roughly the same time but the amount of rain fallen differed significantly (Fig. 5.14) so the correlation was relatively low (R = 0.46) (Fig. 5.15l). Figure 5.17a–b show the relationship between temperature and wind speed for the melt season on both glaciers. The colored dots are wind blowing up glacier, the gray dots are wind blowing down glacier and the black square indicate the theoretical thresholds for glacier winds i.e. temperature above 0°C and wind velocity not higher than 5 m s⁻¹ (Oerlemans, 2010). The general behavior of the gray clusters within the squares is a slight increase in wind speed with increasing temperature. On Rabots glaciär the gray dots keep relatively well within the square whereas on Storglaciären the gray dots are more spread out. Figure 5.17c–d show the relationship between the wind speed and the vertical temperature gradient for the melt
Figure 5.15. Plot of the correlation of climatic factors between Rabots glaciär (y-axis) and Storglaciären (x-axis). Plotted is 15 min data of a) temperature at 0.5 m above the surface, b) temperature at 2 m, c) relative humidity at 0.5 m, d) relative humidity at 2 m, e) vapor pressure at 0.5 m, f) vapor pressure at 2 m, g) albedo, h) total net shortwave radiation i) shortwave incoming radiation, j) total turbulent heat fluxes k) mean wind speed, l) precipitation.

season on both glaciers. There is no clear linear relationship but for wind speeds above 5 m s⁻¹ the gradient was seldom higher than 2°C. Figure 5.17e–f show the relationship between wind speed and wind speed standard deviation for the melt season on both glaciers. It shows that higher winds speeds increase the wind speed standard deviation ($r^2=0.47$ Rabots glaciär and 0.57 on Storglaciären) on both glaciers.

5.2.2. Monthly differences

Figure 5.18 show monthly positive degree days, total incoming radiation and total precipitation for May to August. On Rabots glaciär the month with the highest positive degree days were June and July whereas on Storglaciären the highest values occurred in July and August. On both glaciers the total incoming radiation was highest in May. On Rabots glaciär the decrease is gradual compared to Storglaciären where the incoming total radiation decrease from May to June was great. The amount of total precipitation started relatively low in May and peaks in July on both glaciers. June was the only month when the precipitation was higher on Rabots glaciär (168.3 mm compared to 164.8 mm). However, the amount of hours with rain is higher at Storglaciären for every studied month (Table 5.1).

Figure 5.19 show the monthly correlation between different parameters between
Rabots glaciär and Storglaciären. The temperature correlation between the glaciers were excellent for all investigated months (R = 0.91–0.97). The monthly correlation of relative humidity between the glaciers was considerably higher in May (R = 80) than in August (R = 57). The incoming shortwave radiation correlation between the glacier was highest in May (R = 85) and lowest in July and August (R = 77). In the radiation plots the data follow a bow shaped curve that is very clear in May and becomes more diffuse for each month. The wind speed data correlated badly (R = 23–53). The precipitation shows a great variation in correlation. In May the value is lowest (R = 27) and the highest correlation can be found in June (R = 69).

From June to August the mean vertical temperature gradient (Fig. 5.20a–b) had a constant behavior on both glaciers. The gradient from 0.5 m to 1 m was similar on the glaciers but the gradient from 1 m to 2 m was greater on Storglaciären. The vertical gradient for the melt season was 0.9°C m⁻¹ on Rabots glaciär and 1.0°C m⁻¹ on Storglaciären (Table 5.1).

The behavior of the mean vertical relative humidity gradient (Fig. 5.20c–d) differs significantly between the glaciers. On Rabots glaciär the June to August gradients were almost linear and decreased 3.3–3.7 % from 0.5 m to 2 m. On Storglaciären the June to August mean gradient decreased approximately 5 % from 0.5 m to 1 m and
Figure 5.17. Plots illustrating: the relationship between temperature and wind speed on a) Rabots glaciär and b) Storglaciären; the relationship between vertical temperature gradient as a response to wind speed at c) Rabots glaciär and d) Storglaciären; linear regression with wind speed standard deviation as a response to wind speed at e) Rabots glaciär and f) Storglaciären, for the melt season.

Figure 5.18. Monthly sums of different meteorological parameters. a) positive degree days, b) radiation, c) precipitation.

increased about 2 % from 1 m to 2 m. The gradient from 0.5 m to 2 m is, however very similar (3.3–3.8%).
Figure 5.19. The monthly correlation between temperature (1\textsuperscript{st} row) relative humidity (2\textsuperscript{nd} row) incoming radiation (3\textsuperscript{rd} row) wind speed (4\textsuperscript{th} row) precipitation (5\textsuperscript{th} row) for the investigated months at Rabots glaciär (y-axis) and Storglaciären (x-axis).

The mean vertical vapor pressure gradient (Fig. 5.20e–f) differs greatly. On Rabots glaciär the vapor pressure increased greatly between 0.5 m and 1 m and decreased between 1 m and 2 m. On Storglaciären the vapor pressure hardly changed between 0.5 m and 1 m but increased significantly between 1 m and 2 m.

The amount of clouds changed over the season. On both glaciers May was the month with the highest amount of days with clear or partly cloudy skies. June, July and August were cloudy often with a thick cloud layer. Throughout the season the amount of clear days were very similar between the glaciers but partly cloudy days
Figure 5.20. Vertical temperature at a) Rabots glaciär and b) Storglaciären, relative humidity gradient at c) Rabots glaciär and d) Storglaciären and the vapor pressure gradient at e) Rabots glaciär and f) Storglaciären.

Figure 5.21. Approximation of cloudiness from the shape of the incoming shortwave radiation curve at Rabots glaciär and Storglaciären.

were more common on Storglaciären and thick cloud layers were more common on Rabots glaciär. Note the over 50% of thick cloud layer on Rabots glaciär in August.
5.3. Turbulent energy fluxes and longwave radiation

During the melt season at Rabots glaciär and Storglaciären, the turbulent heat fluxes contributed with an average of 41 W m\(^{-2}\) and 42 W m\(^{-2}\), respectively (Table 2.2) and had the same fluctuations (R = 0.97) (Fig. 5.15). The magnitude of the turbulent heat fluxes are relatively stable in June to August with Storglaciären slightly higher than Rabots glaciär (Fig. 5.22). The relative contribution of the turbulent heat fluxes was greater on Rabots glaciär (16% to 43%) compared to Storglaciären (12% to 32%) (Fig. 5.22). On both glaciers throughout the melt season, the sensible heat was greater than the latent heat. The average sensible heat was 28 W m\(^{-2}\) on Rabots glaciär and 32 W m\(^{-2}\) on Storglaciären glaciers and the average latent heat was 7 W m\(^{-2}\) and 9 W m\(^{-2}\) on Rabots glaciär and Storglaciären, respectively (Table 2.2).

Figure 5.22. Energy measured and calculated on Rabots glaciär (blue) and Storglaciären (orange). a) daily mean of sensible and latent heat, b) monthly means of the sensible heat, latent heat net shortwave radiation and longwave radiation (Storglaciären only) and c) relative contribution of the sensible heat, latent heat and net shortwave radiation.

Figure 5.23 show the seasonal overview of the net longwave radiation on Storglaciären. The incoming longwave radiation seems to have had an inverse behavior to the incoming shortwave radiation and the outgoing longwave was relatively constant from the middle of May. Therefore the net longwave had an inverse behavior
of the shortwave radiation. The constant outgoing longwave radiation was high and the only positive net longwave radiation values occur incoming radiation was high. Figure 5.24 show the relative importance of net shortwave and longwave radiation,

![Figure 5.23. Illustration of daily values of incoming and outgoing longwave radiation and net shortwave radiation on Storglaciären.](image)

sensible and latent heat fluxes on Storglaciären throughout the melt season. Net radiation, sensible and latent heat together contributed on average 123 W m\(^{-2}\) where 81 % (99 W m\(^{-2}\)) came from net radiation, 26 % (32 W m\(^{-2}\)) from sensible heat, 7 % (9 W m\(^{-2}\)) from latent heat and longwave radiation that counteract with -14 % (-18 W m\(^{-2}\)).

![Figure 5.24. Monthly mean temperature, relative humidity and wind speed on Storglaciären within the melt season. May* indicates that the May values are only from 16 May to 31 May, 2013.](image)
5.4. Change in surface height

Figure 5.25 show the cumulative change in glacier surface height for the studied period. Until the middle of May the surface on Rabots glaciär was relatively stable whereas the surface of Storglaciären increases and decreases on a small scale. From the middle of May until late June the surface lowered with roughly the same pattern. In late June the ablation of Rabots glaciär flattened out whereas the ablation on Storglaciären continued. From the middle of July and until the end of the season the surface lowering became more similar again but with a steeper ablation curve on Storglaciären. Over all the correlation of daily water equivalent height loss between the glaciers was 0.57.

![Figure 5.25](image)

*Figure 5.25. Evolution of surface height illustrated as the cumulative ablation on Rabots glaciär (blue) and Storglaciären (orange) for the investigated period.*

The total lowering of the surface (excludes melt of snow fallen within the melt season) was 3.8 m at Rabots glaciär and 4.2 m at Storglaciären (Table 5.1). The water equivalent ablation (includes melt of snow fallen within the melt season) was 2.7 m on Rabots glaciär and 3.2 m on Storglaciären. This means an average ablation of 26 mm d\(^{-1}\) (water equivalent) on Rabots glaciär and 31 mm d\(^{-1}\) (water equivalent) on Storglaciären. In the beginning of the season the change in surface height was ablation of snow and from 28 June for Rabots glaciär and 23 June for Storglaciären the change in surface height was ablation of ice.

![Figure 5.26](image)

*Figure 5.26. Daily water equivalent ablation over the melt season at Rabots glaciär (blue) and Storglaciären (orange).*
Figure 5.26 shows the daily ablation in mm w.e. on both glaciers during the melt season. The ablation fluctuated with roughly the same pattern ($R = 0.57$) on the glaciers but several stand-alone peaks and sinks can be seen. Figure 5.27, illustrates the ablation correlation between the glaciers. The correlation between the ablation of snow is very high ($R = 0.88$) whereas the correlation of ablation of ice ($R = 0.67$) is significantly lower.

**Figure 5.27.** The correlation of daily water equivalent ablation for snow and ice with Rabots glaciär on the y-axis and Storglaciären on the x-axis.

On Rabots glaciär the monthly ablation was evenly distributed over the investigated months compared to the large increase in ablation in July and August on Storglaciären (Table 5.14).

In Figure 5.28 different parameters that theoretically could have influenced the change in surface height were plotted against water equivalent ablation. Temperature and turbulent heat fluxes were the parameters that clearly showed signs of influencing ablation. In the temperature plot for Rabots glaciär, 52% can be explained by the linear regression and 44% for the curve for Storglaciären. The slopes of the correlation lines were similar. Correlating the turbulent heat fluxes to water equivalent ablation the linear regression curve explain 51% of Rabots glaciär ablation and 45% of Storglaciären.
Figure 5.28. Different meteorological parameters as an explanation to daily water equivalent ablation on Rabots glaciär (blue) and Storglaciären (orange) during the melt season. Ablation as a response to: a) temperature at 0.5 m, b) temperature at 2 m, c) relative humidity at 0.5 m, d) relative humidity at 2 m, e) turbulent heat fluxes at 2 m, f) total incoming shortwave radiation, g) wind speed, h) precipitation.
Looking closer at the temperature as an explanation for water equivalent ablation and separating the values for when the change in surface height was ablation of snow or ablation of ice (Fig. 5.29) a difference in the slope of the coefficient could be seen. The change in surface height coefficient for snow on Rabots glaciär was 9.1 mm °C⁻¹, 8.6 mm °C⁻¹ on Storglaciären and for ice 7.5 mm °C⁻¹ and 8.0 mm °C⁻¹, respectively. However, the regression coefficient was lower in comparison to the values for the entire season.

Figure 5.29. Daily ablation for snow (gray) and ice (turquoise) as a response to daily mean temperature for a) Rabots glaciär and b) Storglaciären. Written is the ablation coefficient for snow and ice, respectively and the $r^2$ value connected to the regression line.
6. Discussion

6.1. Meteorological overview

6.1.1. Temperature and humidity

The mean 2 m temperature for the melt season was 4.8°C on Rabots glaciär and 5.6°C on Storglaciären. Previous studies on Storglaciären show similar results. Both a study by Hock and Holmgren (1996) and Sicart et al. (2008) had a mean temperature of 5.4°C (Table 2.2). Studies in Norway (Andreassen et al., 2008; Giesen et al., 2009) show a mean temperatures in the same magnitude (4.9°C to 5.6°C) (Table 2.2). The mean temperature in May to August in Kiruna (Fig. 3.2a) was constantly a few degrees higher which is probable considering that Tarfala is on a significantly higher altitude and on a glacier.

On both glaciers the temperature peaked not long after it stabilized above zero and as the season progressed there was no clear boundary separating the seasonal change between spring and summer (Fig. 5.2 and 5.3). In the middle of August the temperature was relatively stable at a slightly lower temperature which indicate the beginning of fall. In the data set the short time fluctuation is interpreted as the diurnal rhythm. The temperature rises in the daytime and falls in the nighttime due to the daily solar radiation oscillation (Bristow and Campbell, 1984). The temporally longer fluctuation that can be seen throughout the season is likely to be caused by alternating cyclones and anticyclones (Barry, 2008). The positive degree day pattern differs little from the mean temperature due to the small amount of measurements below zero. The difference in temperature within the diurnal rhythm is smaller than the difference in temperature between the cyclones and anticyclones. This demonstrate the importance of the synoptic system as a supplier of energy (Oerlemans, 2010).

The coefficient of variation can be interpret as an indicator of how stable the parameter is within the investigated period. As the 0.5 m temperature sensor is likely to be strongly influence by the glacier surface as well as the atmosphere the sensor measurements was expected to have a higher coefficient of variation compared to the 2 m sensor. However, the diurnal rhythm and the cyclonic fluctuations makes for naturally high variation at all levels.

Mean relative humidity for the melt season was 81.5% on Rabots glaciär and 78.5% on Storglaciären. Studies in Norway (Andreassen et al., 2008; Giesen et al., 2009) show a mean relative humidity in the same magnitude (77.7°C to 80.8°C) (Table 2.2).

Studying the behavior of relative humidity and temperature simultaneously will describe the change in vapor pressure and consequently suggest which processes might be acting at the atmosphere-glacier interface. Theoretically the behavior of the relative humidity curve should inversely follow the temperature curve if no water vapor is added or removed (Barry, 2008). As expected the relative humidity
measurements from 0.5 m and 2 m generally followed the inverse pattern of the temperature curves (Fig. 5.14) with values at 0.5 m slightly higher than at 2 m. However, the differences between the sensor outputs are relatively small and throughout the season the coefficient of variation was low and generally slightly lower at 0.5 m. This can be puzzling considering the large difference and high coefficient of variation recorded for the temperature at both levels. If the relative humidity is stable while the temperature rises it indicates an increase of vapor pressure. However, the relationship is almost exponential and changes in temperature at low temperatures have little effect on relative humidity. For instance, if the air has a temperature of 4°C, and vapor pressure of 670 Pa the relative humidity will be 82.4%. If the temperature increases by 1°C and the vapor pressure is constant the relative humidity will be 76.8%. Consequently, the temperature will have increased 25% but the relative humidity will only have decreased by 7%. This could explain the small difference between the sensor output but also explain why the 0.5 m coefficient of variation for relative humidity was relatively low when the coefficient of variation for temperature was high (Table 5.1). An other explanation of the small difference is the high occurrence of saturated or close to saturated values. The relative humidity can only reach 100% so in high humidity the values at both layers will both reach their individual dew point, condensate and record the same value of relative humidity.

The cooling effect of air caused by sublimation, evaporation and melting will cool the air but at the same time the sublimation and evaporation will add water vapor to the air. This is likely the cause to why, during anticyclonic events within the melt season, the high temperature is relatively stable while the relative humidity slowly increases. When the anticyclones arrived the temperature increased abruptly and consequently the relative humidity decreased. The high temperature and solar radiation would have increased the melt, evaporation and sublimation which gradually added water vapor to the air and consequently increased the relative humidity.

The quick fall of relative humidity often occurring after a longer period of rainfall is probably due to a combination of processes. Since precipitation has been released the air is no longer saturated and during the rain fall moist is constantly evaporating which efficiently cools the air and increases the vapor pressure and consequently induce more precipitation. Followed by an anticyclone the temperature increased significantly making the air unsaturated.

**Vertical gradients**

Due to the processes acting in the atmosphere-glacier interface on a melting glacier (subsection 2.1.3) the general pattern of the vertical temperature gradient was expected. When the temperature was above zero it generally increased with height with the strongest gradient close to the surface (Fig. 5.20). Since the surface of a melting glacier is 0°C the vertical temperature gradient naturally become higher during anticyclones (Oerlemans, 2010). The general appearance of the vertical temperature gradient was similar to the extensive studies by Oerlemans (2010).

During a few days in April, when the weather station was not yet installed on Storglaciären the temperature on Rabots glaciär was well below zero. The daytime temperature at 0.5 m was higher than the temperature at 2 m and at night it became significantly colder at the lower sensor (Fig. 5.2). The 0.5 m temperature fluctuated between −5°C and −20°C whereas the 2 m temperature ranged between −5° and
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−10°C. This could be explained with the sublimation and deposition processes close to the surface (Schmidt and Gluns, 1992). During these days the incoming shortwave radiation were high and the shapes of the daily radiation curves were interpreted as clear sky or partly cloudy (Fig. A.2a1). As the surface was fed with high solar radiation, sublimation slowly began to cool the air above and when the 0.5 m temperature fell below the 2 m temperature the warmer air above enhanced the processes and the temperature dropped further. As the incoming radiation declined deposition began to heat the 0.5 m air well before the incoming radiation returned to contribute to the heating. This can be supported by the behavior of the relative humidity which theoretically should inversely follow the temperature curve if no water vapor is added or removed. The relative humidity at 2 m inversely followed the temperature curve at 2 m but at 0.5 m the humidity was relatively constant and peaks and sinks simultaneously with the temperature. This indicates an addition of water vapor to the air at high temperature and a removal at low temperature.

6.1.2. Radiation

Similar to the temperature data the incoming solar radiation show a clear diurnal rhythm (Bristow and Campbell, 1984) and a temporally longer fluctuation that can be seen throughout the season and is likely to be caused by cyclones and anticyclones (Barry, 2008). In June and a large part of July the incoming radiation does not reach zero due to the of midnight sun.

Incoming shortwave radiation in the energy balance studies presented in table 2.2 show a variation from 133 Wm$^{-2}$ to 284 Wm$^{-2}$ where the lowest presented value is from Storglaciären (Sicart et al., 2008). In Hock and Holmgren (2005) incoming radiation values from Storglaciären was 147 Wm$^{-2}$ in 1993 and 169 Wm$^{-2}$ in 1994. The mean incoming radiation in May to August in Kiruna (Fig. 3.2b) followed similar patterns. Most interesting to see is the drop in radiation in June that is visible in all three locations.

Since an overcast day in summer can contain more energy than a clear day in winter the sum of shortwave radiation presented is difficult to interpret. When studying the interpretation of the shape of the radiation curve (Fig. 4.2), May was least cloudy, June and July was equally cloudy and August slightly less cloudy (Fig. 5.21). Considering that the summer solstice is 21 June on the northern hemisphere the decrease of the sum of the monthly shortwave radiation on both glaciers (Fig. 5.18) are likely to be caused by length of the days as well as cloudiness. In contrast to the temperature fluctuation the difference within the diurnal rhythm is significantly greater than the effects caused by synoptic pressure gradients which demonstrate the importance of incoming radiation as a constant supplier of energy.

As expected the coefficient of variation is very high since the mean values represent both daily fluctuations and fluctuations caused by changes in the synoptic system.

Albedo

On both glaciers high albedo in the beginning of the season indicates that relatively fresh snow covered the surfaces. During this time a few peaks suggest fresh snowfall. In the middle of May the temperature briefly reached above zero changing the surface texture and lowering the albedo. In late June a large drop in in the albedo curve on Rabots glaciär indicate relatively quickly change from snow to ice. On Storglaciären
the albedo drop arrives in the middle of June but is less sharp indicating a period of slush. However, from late June the ice is exposed on both glaciers with the exception of a few snowfall throughout the season that can be seen as brief peaks in albedo. Smaller changes in albedo was likely to be caused by changes in the roughness of the ice surface. Rabots glaciär constantly has a minimum of 0.1 higher albedo for both the snow and ice covered surface. Comparing the snow and ice albedo to other studies both glaciers (Table 2.1) have likely values. The differences could therefore be caused by diverging surface properties but it is also likely to be an effect of the different sensitivity of the sensors.

6.1.3. Wind direction and speed

On a valley glacier in summer, the air flow can be of different origin. Stronger synoptic winds that behaves differently depending on the direction, size and shape of the wind and the mountain, valley- and mountain winds with a diurnal flow and glacier winds with low velocity and is of katabatic origin (Barry, 2008). Figure 5.17a–b indicates an increase of velocity with temperature for low velocity winds in summer and is a feature of a glacier wind (Oerlemans, 2010). The tails reaching up to higher wind velocity was likely of synoptic origin. Figure 5.17c–f show that the temperature gradient is higher, and wind speed standard deviation is low for low velocity winds. These stable qualities are also typical features for a glacier wind (Oerlemans, 2010). The presence of a glacier wind can bee seen more clearly in figure 5.9 and 5.10 which show the velocity and direction of the winds from April to August and figure 6.1 that show the spatial extent of the wind directions.

On Rabots glaciär the wind mainly came from either the two upper cirques or from the foot of the glacier. The wind speed show that the cirque winds were seldom higher than 4 m s$^{-1}$ meanwhile the wind from the foot seldom was calmer than 2 m s$^{-1}$. This indicates that the wind from the cirques were of katabatic origin whereas the wind flowing up valley was either a valley wind or part of a greater synoptic system. The wind coming from the middle cirque could also be explained or enhanced by a wind movement across the glacier caused by intense heating of the south-facing slope consequently making the light air lift creating a low pressure zone on the sunny side forcing air to flow in from higher pressure areas (Barry, 2008). The air coming in from the foot of the glacier lack warmer and drier air whereas air from the cirques at times contain both. Looking at the topography of the valley this could be explained with the valley wind cooling and consequently gets higher relative humidity as it flows up valley and the cirque winds heating and drying as it descends the steep cirques. From the upper cirque there was less slope but a longer distance to travel to the weather station. The middle slope was near but the decent is around 400 m. On Storglaciar, a significant amount of the wind came from up glacier. Looking at the topography (Fig. 6.1) the prevailing westerly synoptic wind and a glacier wind should both flow down glacier. Considering the upper cirque was facing south a thermal induced wind was also likely to flow down glacier. A few percent of the winds came from other directions but were seldom higher than 4 m s$^{-1}$ and contained no warm or dry air. The easterly wind (up glacier) made up a large fraction (11 %) of the diverging directions. This is likely to originate from the valley wind flowing up the Tarfala valley and side track up the glacier.

The mean wind speed for the melt season was 2.5 m s$^{-1}$ on Rabots glaciär and
Figure 6.1. Wind direction that make up more than 10% of the directions for the studied period at; a) Rabots glaciär (1 April to 5 September 2013), b) Storglaciären (18 April to 5 September 2013), c) Tarfala Research Station (2013). The size of the arrows indicate the fraction of the wind direction.

3.0 m s\(^{-1}\) on Storglaciären. Table 5.1 show the mean wind speed for different melt season studies around the globe. The wind speed ranges from 2.1 m s\(^{-1}\) on Worthington Glacier, USA (Streten and Wendler, 1968) to 6.0 m s\(^{-1}\) on Midtdalsbreen, Norway (Giesen et al., 2009) but is most common just above 3 m s\(^{-1}\). A previous study on Storglaciären presents a mean value of 2.5 m s\(^{-1}\) but this is measured only during one month in July/August (Hock and Holmgren, 1996).

In figure 6.1 the main wind direction from an automatic weather station situated at Tarfala Research Station is added. Both a mountain wind and valley wind can be seen. However, the winds with highest wind speed are northerly likely originating from west but flow through the valley close to Rabots glaciär and squeezes through the pass down to the Tarfala valley.

6.1.4. Precipitation

The precipitation data fits relatively well into what earlier has been interpreted as cyclones. However, some high peaks did not fit this pattern. At these events the incoming shortwave radiation temporarily dropped from high values to low values within half an hour (Fig. A.2). This was interpret as *cumulonimbus clouds* which are usually dense, penetrate little radiation and has the potential to release great amount of precipitation. Figure 5.11 indicate that the precipitation on Rabots glaciär came
from all directions. On Storglaciären, where the wind direction was more homoge-
neous it was surprizing to see that the wind directions bringing precipitation was not as
dominant. The fraction of easterly winds increased significantly relative to the
general dominant wind directions. This indicates that humid air brought by the
valley wind from the Tarfala valley, cooled when it came in contact with the glacier
and released precipitation.

Precipitation in Kiruna differs from the glaciers. The values on the glaciers are
significantly higher but follow a similar pattern. Considering that the prevailing
wind in Northern Sweden is westerly and Kiruna is situated east of the Kebnekaise
massif it is possible that Kiruna is in rain shadow of the mountains.

However, the precipitation is likely to originate from the synoptic system and
released from a higher air layer. The wind direction recorded would then be the
katabatic flow working relatively independent from the synoptic system. It is there-
fore unlikely that the cardinal origin of precipitation is unknown.

6.2. Meteorological differences

The high temperature correlation between the glaciers indicate that the reoccurring
fluctuations over the season is controlled by the synoptic system. That the correla-
tion was higher at 2 m is therefore expected since the top sensor was more influenced
by the synoptic system and less by the processes at the atmosphere-glacier interface.
The slightly higher mean temperature on Storglaciären could be explained with the
categorized shape of the daily radiation curves which shows a generally thicker cloud
layer over Rabots glaciär.

The higher temperature gradient at Storglaciären could indicate less turbulence
but a higher temperature at 2 m will often naturally lead to a higher vertical gradient
since the surface temperature is 0 °C. However, the homogenous wind direction at
Storglaciären is likely to cause less turbulence but the higher wind speed standard
deviation seen on the glacier likely causes more turbulence.

The correlation of relative humidity between the glaciers was not as strong which
could be explained with the high correlation of vapor pressure. The correlation of
relative humidity and vapor pressure between the glaciers are greater at 2 m again
indicating that the higher sensor was more influenced by the synoptic system.

The change of relative humidity between 0.5 m and 2 m is expected considering
that the increase in temperature at the higher level should lead to lower relative
humidity. The vapor pressure gradient at Rabots glaciär behaves as expected. The
1 m layer contain warmer air than at 0.5 m and can therefore hold more vapor and is
close enough to the surface to receive vapor from ongoing processes. The air at 2 m
is warmer still but further from the atmosphere-glacier processes and consequently
less affected by them. At Storglaciären the warmer air at 1 m does not hold a
significantly higher amount of vapor than at 0.5 m and instead the great change
occur at 2 m. The small change in vapor pressure at the lowest level at Storglaciären
indicate little addition or more removal of water vapor hence less evaporation or more
condensation.

Considering that the shortwave radiation is measured at different range of wave-
lengths comparing shortwave radiation between the glaciers was difficult. The peak
of the intensity of incoming radiation on earth is covered by the 300 nm to 1000 nm
sensor and a huge amount at radiation above 1000 nm is absorbed or reflected by wa-
ter vapor before it reaches the surface (Thekaekara, 1965). The amount of 1000 nm to 2800 nm radiation reaching the surface should therefore be relatively small and likely the difference between the sensors would be greater in overcast conditions compared to clear conditions. The values on Storglaciären are however significantly greater than on Rabots glaciär and studying the distribution of the categorized shape of the daily radiation curve this is probably not all due to the measured wavelength range but to actual less incoming shortwave radiation due to more or thicker clouds and shading on Rabots glaciär. This is supported by the relatively low correlation between incoming radiation compared to the temperature and vapor pressure. The shortwave radiation sensor produce less noise (Fig. A.2) and the different range of the sensors could affect the albedo values. In a comparison between two Norwegian glaciers Giesen et al. (2009) concluded that the lower mean albedo on one glacier was caused by generally less winter precipitation on the site which lead to an earlier exposure of ice. At the site for the weather stations the mean snow depth for the years 2005–2011 was 1.9 m at Rabots glaciär and 1.4 m at Storglaciären (Data obtained from Bolin Centre, http://bolin.su.se/data/tarfala/glaciers.php, 9 Jun., 2014). This is likely the reason why the ice was exposed earlier on Storglaciären than on Rabots glaciär. However, the difference in albedo is too great for this to be the singular reason for the significantly higher albedo value. The difference in albedo likely reflect the difference in surface roughness. In the albedo correlation (Fig. 5.15g) the three clusters make up the data points for snow, old snow and ice and the noise between old noise and ice is interpret as slush. The bow shaped pattern in the correlation of incoming shortwave radiation is likely the daily cycle of the sun. The pattern is more clear in May when clear skies are common and less distinct in August when the conditions are more cloudy and the sun is low.

The most striking difference in the data set is the wind directions that at times blow in the opposite direction from each other. The difference between the wind speeds is likely explained by how the glaciers are situated in the massif in relation to the prevailing westerly wind. On Storglaciären the prevailing wind and the katabatic flow will work together whereas on Rabots glaciär they counteract. The valley where Rabots glaciär is resting is larger compared to Storglaciären which is a smaller valley almost perpendicular to the greater Tarfala valley. Therefore it is likely that a collision with a greater valley wind system seldom occur on Storglaciären. Further, the westerly wind need to climb up Rabots glaciär loosing energy whereas on Storglaciären the wind was flowing with gravity.

In more extensive study of the difference in micro-climate on two glaciers, 120 km apart, in Southern Norway, Giesen et al. (2009) conclude that the wind at Storbreen had a katabatic flow down glacier and Midtdalsbreen had a wind direction that indicate strong influences from the synoptic system. The different origin of air flow on these glaciers show great difference in wind speed which support the theory of katabatic and synoptic flow. The difference in wind speed at Rabots glaciär and Storglaciären are not as great and katabatic flow and synoptic winds is likely present on both glaciers.

The two conditions for wind gusts to occur, relevant to this area, is gusts developed when the wind passes over or around a barrier and gusts that occur due to surface turbulence. The latter is smaller in space, time and in magnitude. As seen in figure 5.17 there was a moderate relationship between wind speed and wind gusts. However, considering high speed winds originate from generally the same direction
on the glaciers the occurrence of wind gusts might be due to the originating wind properties and not the wind speed itself. Studying the morphology in the up-wind main direction for strong winds the wind on Storglaciären ought to be influenced by more varying friction from surrounding topography. This could explain the higher values of wind speed standard deviation interpret as occurrence of wind gusts. However, Rabots, glaciär should have a generally greater occurrence of small scale wind gusts considering greater turbulence at the surface.

The greater amounts of precipitation at Storglaciären could be puzzling considering Rabots glaciär has less incoming radiation and generally more cloudiness. However, the amount of precipitation is 32% greater on Storglaciären but the amount of 15-min recordings with precipitation are only 15 15% greater. The relatively low correlation between precipitation on the glaciers can indicate that it does not always rain on both glaciers at the same time but it can also indicate that the amount of rainfall does not correlate. When the precipitation originates from east on both glaciers, it is likely that there could be a rain shadow effect on Rabots glaciär. Incoming air with high humidity needs to rise at least 300 m to get over the peak.

In a study by Jansson et al. (2007) were the possibilities to couple chemical signals in winter accumulation on Storglaciären to atmospheric climatology was investigated, they conclude that winter precipitation on the glaciers originates from several different directions. Similar results can be seen in the summer precipitation data on both Rabots glaciär and Storglaciären. On Storglaciären it is especially striking considering the general lack of easterly winds when there is no precipitation.

6.3. Energy fluxes

The turbulent heat fluxes is the factor that correlated best between the glaciers. Compared to Storglaciären the turbulent heat fluxes made up a larger portion of the combined latent heat, sensible heat and shortwave radiation on Rabots glaciär. This is however, likely due to the different range of the radiation the sensors measures. Both the latent heat and the sensible heat are greater on Storglaciären and considering the theory of less vapor present the larger amount of latent heat would suggest more condensation on Storglaciären. The negative latent heat seen in May is relatively equal on the glacier but is slightly larger on Rabots glaciär. These negative values are likely due to sublimation. However, all turbulent flux values are computed using a simple method and the margin of error are expected to be large and is presented only to get a rough estimation of the magnitude of the parameters affecting ablation.

Table 2.2 show measurements for incoming shortwave radiation and calculated values for sensible and latent heat fluxes for different studies around the globe. The measured mean incoming shortwave radiation for the melt season on Rabots glaciär (149 W m⁻²) and Storglaciären (181 W m⁻²) fell within the range of the global values. So did the calculated sensible and latent heat fluxes (32 W m⁻² and 9 W m⁻² on Rabots glaciär and 32 W m⁻² and 10 W m⁻² on Storglaciären). The turbulent heat fluxes differs slightly but not significantly from previous studies on Storglaciären (Hock and Holmgren, 1996; Hock and Holmgren, 2005; Sicart et al., 2008) where the sensible heat fluxes ranges between 33 W m⁻² to 44 W m⁻² and the latent heat fluxes ranges between −4 W m⁻² to 14 W m⁻². The incoming shortwave radiation and net radiation is higher than the average of the studies from Storglaciären but
is similar to the results from Hock and Holmgren (1996). This study is remarkably similar in the relative contribution of the parameters which is also the case for both studies on Midtdalsbreen, Norway (Giesen et al., 2008; 2009).

Comparing results between different studies are difficult since most studies are done at different length of time, at different time within the melt season and at different years. Studies show the importance of weather for the relative importance of the parameters. Clear skies are expected to increase the importance of net radiation and overcast conditions increase the importance of the turbulent heat fluxes (Hock, 2005). Surface properties (due to albedo), altitude (due to vertical temperature lapse rates), instrument accuracy and computing method will also change the results. Giesen et al. (2009) noticed a significant decrease (13 % and 18 %) in net radiation when including clear sky nights when the temperature dropped below zero. Since longwave radiation was not measured on Rabots glaciär the relative importance of the parameters are difficult to compare to other studies. Figure 5.22c show the relative importance of net shortwave radiation, sensible heat flux and latent heat flux alone. Since the net shortwave radiation was significantly smaller on Rabots glaciär and the turbulent fluxes was similar between the glaciers the relative importance of the turbulent fluxes are higher on Rabots glaciär. However since Rabots glaciär seems to be cloudier than Storglaciären the longwave radiation is likely to be higher than the longwave radiation measured at Storglaciären. If so, it would lessen the significant difference in energy suggested by the higher net shortwave radiation on Storglaciären (101 W m$^{-2}$ versus 56 W m$^{-2}$). Giesen et al. (2009) concludes that the high relative importance of net radiation at Storbreen, Norway is caused by it being less maritime than several comparative studies and had lower albedo. Considering Rabots glaciär and Storglaciären should be influenced by the same synoptic climate forcing the explanation in this study is more likely micro-climate and albedo. This study did not exclude short periods temperature below zero within the melt season which might have decreased the net radiation and therefore increased the relative importance of the turbulent heat fluxes (Giesen et al., 2009).

### 6.4. Change in surface height

The weeks following the installation of the stations both glaciers have relatively stable surfaces except for a few minor snowfall on Storglaciären. After a large snowfall on both glaciers in the middle of May the temperature briefly reached above zero which changed the surface texture reflected in the albedo curve (Fig 4.3). After this the surface melted and compressed steadily on both glaciers until a short period in the middle of June on Storglaciären and in late June on Rabots glaciär when slush covered the surface until the ice was exposed. This is interpreted from the individual albedo curves that made a great drop (from 0.7 to 0.4 and 0.6 to 0.3, respectively) until stabilizing on a lower value. Until early August the surface lowering on the glaciers are similar with Storglaciären constantly lowering slightly more. From early August the curves behave similarly but the surface on Storglaciären lowers at a higher pace. August is the month where both temperature, radiation and turbulent fluxes are significantly higher at Storglaciären which is likely to explain the rapid ablation (Hock, 2005). That Storglaciären and Rabots glaciär correlate less during ice ablation than snow ablation is due to greater difference in energy available. This is likely caused by the significantly higher albedo (0.2 higher in June and July) at
Storglaciären in the peak of the melt season. Giesen et al. (2009) witnessed the same behavior when comparing ablation on Storbreen and Midtdalsbreen. Midtdalsbreen had 0.05 lower albedo than Storbreen which was mainly due to the AWS site on Midtdalsbreen being snow free 11–35 days earlier than on Storbreen. These glaciers also showed the same behavior in the albedo curve. The glaciers exposing the ice first (Midtdalsbreen and Storglaciären) had a longer period of what can be interpret as slush whereas Storbreen and Rabots glaciär goes from snow to ice relatively quickly. Giesen et al. (2009) explains the earlier exposure of ice on Storglaciären mainly due to a deeper snow accumulation at the AWS site on Storbreen. Considering Storglaciären having approximately 0.5 m less snow than Rabots glaciär at the AWS sites in 2005–2011 the same explanation is likely.

The higher ablation rate on Storglaciären compared to Rabots glaciär is caused by a greater amount of energy available for melt from net shortwave radiation and the turbulent heat fluxes. The ablation rate on Storglaciären is similar to the ablation rates from Hock and Holmgren (1996) (32 mm d\(^{-1}\) w.e.) and Hock and Holmgren (2005) (30 mm d\(^{-1}\) w.e. in 1993 and 34 mm d\(^{-1}\) w.e. in 1994). The few ablation rates presented from the chosen surface energy balance studies (Table 2.2) show a wider range (20 mm d\(^{-1}\) to 70 mm d\(^{-1}\) w.e.). Generally the ablation rates from Greenland (Braithwaite and Olesen, 1990; van de Wal et al., 2005) and South georgia (Hogg et al., 1982) were lower and the ablation rates at temperate glaciers in USA (Streten and Wendler, 1968) and New Zealand (Hay and Fitzharris, 1988) was higher. Like Konya et al. (2004) Storglaciären and Rabots glaciär had a lower ablation rate for snow than for ice. This is likely due to the significantly lower albedo for ice that indicate less absorption of shortwave radiation.

The glaciers have similar slope on the regression line for the relationship between temperature and ablation indicating that a similar response in ablation occur when the temperature rises. Both glaciers have a lower ablation coefficient for snow than ice. This is likely cause by the thermal conductivity of snow (0.05 W m\(^{-1}\) K\(^{-1}\) to 0.25 W m\(^{-1}\) K\(^{-1}\)) that is significantly lower than for ice (2.18 W m\(^{-1}\) K\(^{-1}\)) and the high albedo of snow that repress the incoming shortwave radiation. Both glaciers show ablation when the temperature is below zero. This is likely due to of compression of snow and, in a smaller magnitude, sublimation. The great amount of melt seen in July and August on Storglaciären can be coupled to the higher amount of thick clouds on Rabots glaciär but also in the low correlation of the different parameters between the glaciers in July and August.

Giesen et al. (2009) compared the micro-climate on two glaciers in Norway. The study showed a remarkable similarity in the yearly mean incoming and mean net radiation. In 1942–1948 Wallén (1949) did an extensive study on the relationship between meteorological parameters and ablation on Kårsa glaciären, situated approximately 100 km north of the Kebnekaise massif. The climate on the glacier varied considerably between the years and Wallén concluded that the type of weather event was more important than the time for the event. Both the results from Giesen et al. (2008) and Wallén (1949) could likely be the case if this study was continued over several years. The variations in the radiation cycle is small compared to the diurnal rhythm but for temperature the diurnal rhythm is small compared to the seasonal fluctuation.

To properly be able to compare the energy parameters between the glaciers the same method and setting is needed. The relative importance of heat from rain
and ground heat flux is expected to be small (Wheler, 2009) but to get a proper estimation of the differences these parameters needs to be included. Considering the significance of net radiation for the surface energy balance (e.g. Streten and Wendler, 1968; Hogg et al., 1982; Hay and Fitzharris, 1988; Braithwaite and Olesen, 1990; Munro, 1990; Hock and Holmgren, 1996; Konya et al., 2004; Hock and Holmgren, 2005; van de Wal et al., 2005; Andreassen et al., 2008; Giesen et al., 2008; Sicart et al., 2008; Giesen et al., 2009 ) it would be necessary for further studies to have the same radiation sensors.

In the continuous winter mass balance measurements on glaciers in Northern Sweden it has been observed that Storglaciären has the highest winter mass balance in the Kebnekaise massif (Holmlund and Jansson, 1999). Considering this it would be extremely interesting to expand the study to include the accumulation season and investigate differences in micro-climate between Rabots glaciär and Storglaciären for several years.

6.5. Source of error

Table 4.1 lists the range and the accuracy of the sensors used for measuring the different meteorological parameters. In April to September most meteorological parameters should be within the range for the different sensors. Theoretically the temperature on the glaciers could reach below \(-40 ^\circ C\) in winter but considering that the Swedish minimum temperature record in April is \(-36.5 ^\circ C\) (Data obtained from SMHI, http://www.smhi.se/klimatdata/meteorologi/2.1240, 9 Sep., 2014) and was

![Figure 6.2. Plots for investigating error in precipitation measurements. Wind speed plotted against precipitation on a) Rabots glaciär and b) Storglaciären and incoming shortwave radiation plotted against precipitation on c) Rabots glaciär and d) Storglaciären.](image-url)
recorded in 1916 this is highly unlikely. The wind speed record for April occurred in Tarfala in 2007 but is 50 m s\(^{-1}\) which is well within the sensor range.

When redrilling the sonic ranger stake on Rabots glaciär in July it was set approximately 0.4 m above ground which was below the range for the SR50A. However, since these values appear correct and only averages are used, four days with measurements below the threshold were used.

The accuracy of the sensors were theoretically lessened by the use of averages over relatively long time periods. The radiation sensors had the largest error. This error was likely enlarged due to the sensors probably not always being perpendicular to the constantly changing glacier surface. These errors are difficult to estimate without measuring the angle of the mast throughout the melt season (Giesen et al., 2009).

Precipitation has a relatively poor accuracy. However, especially on Storglaciären some precipitation values seems extremely high. These values were found especially during high wind velocities (Fig. 6.2a–b) and the question if turbulence inside of the catchment bucket could make the tipping bucket tip consequently measuring faulty precipitation. According to SMHI (SMHI, http://www.smhi.se/kunskapsbanken/meteorologi/vaderspraket-1.3847, 9 Sep., 2014) the definition of a shower is rain that starts suddenly, are temporally short and covers a small area. A moderate shower lasts 10 min to 20 min and produces 1 mm to 10 mm of precipitation. Shower falls from *Cumulonimbus clouds* instead of regular rain clouds (*Nimbostratus*) which should result in a very low incoming shortwave radiation. Figure 6.2c–d does show that high precipitation occur when the incoming shortwave radiation is low and the highest precipitation measured in the 15 min -interval (5.5 mm on Rabots glaciär and 8.7 mm on Storglaciären) is within the range of a moderate shower. The high values are therefore difficult to discard. However, the measurements collected by the tipping bucket should be treated with caution. Several studies (e.g. Habib et al., 2001; Molini et al., 2001; Ciach, 2003; Upton and Rahimi, 2003; Chvila et al., 2005) deal with great errors using a tipping bucket. The most common errors were errors when measuring rain intensity at low precipitation on a short timescale and observed effects from high wind velocity was usually an underestimation of precipitation. However, Upton and Rahimi (2003)

My lack of knowledge of the glacier surface properties made it difficult to approximate the stability corrections and roughness lengths needed to make a more complex calculation of the turbulent fluxes. However, the simple formula used is developed for turbulent heat flux processes when a glacier wind is present. Consequently, the error in the turbulent heat flux calculations were believed to be larger at higher wind velocities, likely by underestimating the fluxes, and therefore larger on Storglaciären than on Rabots glaciär.

There are relatively large uncertainties in the surface height data. The margin of error for the sensor is 10 mm which corresponds to almost 50 % of the daily mean water equivalent ablation (24 mm for Rabots glaciär and 28 mm on Storglaciären). The stake at Storglaciären was unstable for a time and even though the measurements has been cleaned up noise is still likely to be present. The water equivalent ablation is only an approximation. When the surface contains snow there will be compression. The compression will overestimate ablation at one hand and underestimate it when considering water equivalent ablation due to heighten density. The density of snow, ice and slush is estimated and within the categories there will be great difference. Minor snowfall can fail to heighten the mean midday albedo enough
to be categorized as snow and snowfall and therefore overestimate water equivalent ablation. When the snowfall does heighten the albedo sufficiently it might melt away the new snow but also melt ice which will underestimate water equivalent ablation.
7. Conclusion

The results for the investigated period shows that there was a difference in microclimate between Rabots glaciär and Storglaciären. Generally Storglaciären had slightly warmer and drier air, had less or a thinner cloud layer but more precipitation. On both glaciers a glacier wind is dominant but high wind velocities were common especially on Storglaciären indicating a larger influence by the synoptic system. The good correlation for temperature and vapor pressure between the glaciers indicate that both glaciers are strongly affected by the synoptic system.

The meteorological parameters have similar effect on the ablation on the glaciers. Temperature, vapor pressure and the turbulent heat fluxes are the only meteorological parameters that suggest a linear affect on ablation.

Net shortwave radiation contribute with the most energy for ablation but decrease in relative importance as the temperature increases. Shortwave radiation, sensible and latent heat contribute with a total $184\,\text{W}\,\text{m}^{-2}$ on Rabots glaciär and $222\,\text{W}\,\text{m}^{-2}$ on Storglaciären. Rabots glaciär seem to have a significantly greater relative importance of the turbulent heat fluxes than Storglaciären. However this could be an effect of the different sensors used for measuring shortwave radiation.

Even though there are differences in micro-climate between the glaciers the differences are not great. But, using simply the ablation for Storglaciären to estimate ablation on Rabots glaciär would over estimate the ablation with 0.5 m w.e. which is a fair amount considering the proximity of the glaciers.

It would be very interesting to continue this study and investigate the annual differences. However, to properly study the difference in micro-climate and the affect it has on ablation changes in the setup would be necessary. It is essential to have identical radiation sensors and to include the accumulation period. It would also be useful to study the surface properties and and if possible measure temperature, relative humidity and wind speed on a few more spots on the glacier to get a spatial perspective of the differences.
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A. Appendix

A.1. Data collection

A.1.1. Table of the data recorded

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<th>Data type (Unit)</th>
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<td>Average (W m(^{-2}))</td>
<td>-/-</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td></td>
<td>Totalize (kJm(^{2}))</td>
<td>-/-</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td>Incoming LW Radiation</td>
<td>Average (W m(^{-2}))</td>
<td>16/-</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td></td>
<td>Totalize (W m(^{-2}))</td>
<td>17/-</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td>Outgoing LW Radiation</td>
<td>Average (W m(^{-2}))</td>
<td>18/-</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td></td>
<td>Totalize (W m(^{-2}))</td>
<td>19/-</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td>Albedo</td>
<td>Average (%)</td>
<td>20/-</td>
<td>( C_{12/-} C_{14/-} )^{[3]}</td>
</tr>
<tr>
<td>Total Incoming Radiation</td>
<td>Average (W m(^{-2}))</td>
<td>21/-</td>
<td>( C_{12/-} + C_{16/-} )</td>
</tr>
<tr>
<td>Total Reflected Radiation</td>
<td>Average (W m(^{-2}))</td>
<td>22/-</td>
<td>( C_{14/-} + C_{18/-} )</td>
</tr>
<tr>
<td>Total Net Radiation</td>
<td>Average (W m(^{-2}))</td>
<td>23/-</td>
<td>( (C_{12/-} + C_{16/-}) - (C_{14/-} + C_{18/-}) )</td>
</tr>
<tr>
<td>LW in Co</td>
<td>Average (W m(^{-2}))</td>
<td>24/-</td>
<td>( \left( \frac{C_{12}}{T} + 5.6710^{-8}(T+273.15) \right)^4 )^{[4]}</td>
</tr>
<tr>
<td>LW out Co</td>
<td>Average (W m(^{-2}))</td>
<td>25/-</td>
<td>( \left( \frac{C_{18}}{T} + 5.6710^{-8}(T+273.15) \right)^4 )^{[4]}</td>
</tr>
<tr>
<td>Wind Speed</td>
<td>(m s(^{-1}))</td>
<td>26/16</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td>Wind Direction</td>
<td>(°)</td>
<td>27/17</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td>Wind Direction (sd)</td>
<td>(°)</td>
<td>28/18</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td>Maximum Wind Speed</td>
<td>(m s(^{-1}))</td>
<td>29/19</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td>Minimum Wind Speed</td>
<td>(m s(^{-1}))</td>
<td>30/20</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td>Wind Speed (sd)</td>
<td>—</td>
<td>31/21</td>
<td>10^6 sec (15^th min)</td>
</tr>
<tr>
<td>Precipitation</td>
<td>Totalize (mm)</td>
<td>32/22</td>
<td>every 0.1 mm (15^th min)</td>
</tr>
<tr>
<td>Raw Distance</td>
<td>(m)</td>
<td>33/23</td>
<td>15^th min (15^th min)</td>
</tr>
<tr>
<td>Signal Quality</td>
<td>—</td>
<td>34/24</td>
<td>15^th min (15^th min)</td>
</tr>
<tr>
<td>Temp. Corr. Distance</td>
<td>(m)</td>
<td>35/25</td>
<td>( \sqrt{\frac{C_{12/3} + C_{15/8} + k}{k}} )^{[5]}</td>
</tr>
</tbody>
</table>

\(^{[1]}\) Column number in logger output table, \(^{[2]}\) 0.5 m, 1 m and 2 m from the glacier surface, \(^{[3]}\) According to logger program, not used, \(^{[4]}\) \( E \), constant given by instrument manufacturer, \(^{[5]}\) \( k \), constant given by instrument manufacturer
A.1.2. Logger program

Rabots glaciär logger program, version 1

'CRI000
'peter.jansson@2013-01-23

'Declare Variables and Units
Public BattV
Public PTemp_C
Public AirT05
Public RH05
Public AirT1
Public RH1
Public AirT2
Public RH2
Public ShkWDN
Public ShMJDN
Public ShkWUP
Public ShMJUP
Public WS_ms
Public WindDir
Public Rain_mm
Public SR50(2)
Alias SR50(1)=RawDist
Alias SR50(2)=SignalQuality
Public TempCorr
Public TempCorrDist

'Define Data Tables
DataTable(Table1, True, -1)
  DataInterval (0, 15, Min, 10)
  Sample (1, AirT05, FP2)
  Sample (1, RH05, FP2)
  Sample (1, AirT1, FP2)
  Sample (1, RH1, FP2)
  Sample (1, AirT2, FP2)
  Sample (1, RH2, FP2)
  Average (1, AirT05, FP2, False)
  Average (1, RH05, FP2, False)
  Average (1, AirT1, FP2, False)
  Average (1, RH1, FP2, False)
  Totalize (1, SlrkWDN, IEEE4, False)
  Average (1, SlrkWUP, FP2, False)
  Totalize (1, SlrMJDN, IEEE4, False)
  WindVector (1, WS_ms, WindDir, FP2, False, 0, 0, 0)
  FieldNames ("WS_ms, WindDir, FP2")
  Maximum (1, WS_ms, FP2, False, False)
  Minimum (1, WS_ms, FP2, False, False)
  StdDev (1, WS_ms, FP2, False)
  Totalize (1, WindDir, FP2, False)
  Sample (1, RawDist, FP2)
  Sample (1, SignalQuality, FP2)
  Sample (1, TempCorrDist, FP2)
EndTable

DataTable(Table2, True, -1)
  DataInterval (0, 1440, Min, 10)
  Minimum (1, BattV, FP2, False, False)
EndTable

'Main Program
BeginProg
'Main Scan
Scan(10, Sec, 1, 0)
'Default Data logger Battery Voltage measurement "BattV"
'Battery (BattV)
'Default Wiring Panel Temperature measurement "PTemp_C"
'PanelTemp (PTemp_C, 60Hz)
'PortSet (9, 1): Turn on SW, 12V
'Delay (0.3, Sec)
'HC283 Temperature & Relative Humidity Sensor measurements "AirT05" and "RH05"
'Volts (AirT05, 1, mV2500, 1, 0.0, 60Hz, 0.1, -40)
Volts (RH05, 1, mV2500, 2, 0.0, 60Hz, 0.1, 0.0)
'HC283 Temperature & Relative Humidity Sensor measurements "AirT1" and "RH1"
'Volts (AirT1, 1, mV2500, 3, 0.0, 60Hz, 0.1, -40)
Volts (RH1, 1, mV2500, 4, 0.0, 60Hz, 0.1, 0.0)
'HC283 Temperature & Relative Humidity Sensor measurements "AirT2" and "RH2"
'Volts (AirT2, 1, mV2500, 5, 0.0, 60Hz, 0.1, -40)
'EndScan
EndProg
VoltsSe(RH2.1,mV250,6,0.0,60Hz,0.1,0)
If RH05>100 AND RH05<103 Then RH05=100
If RH1>100 AND RH1<103 Then RH1=100
If RH2>100 AND RH2<103 Then RH2=100
'SR50 on SDI=12
If ITime(0,15,Min) Then
SD12Recorder(SR50(),1.0,"MI",1.0)
TempCorr=(AirT05+AirT1)/2
TempCorrDist=RawDist*(SQR((TempCorr+273.15)/273.15))
EndIf
PortSet(9,0) 'Turn off SW 12V
'CSE300 Pyranometer measurements 'SlrMJDN' and 'SlrkWDN'
VoltsSe(SlrWDN,1,mV250,7,1.0,60Hz,1,0)
If SlrkWDN<0 Then SlrkWDN=0
SlrkWDN=SlrkWDN+5
'CSE300 Pyranometer measurements 'SlrMJUP' and 'SlrkWUP'
VoltsSe(SlrWUP,1,mV250,8,1.0,60Hz,1,0)
If SlrkWUP<0 Then SlrkWUP=0
SlrkWUP=SlrkWUP+5
'S5103 Wind Speed & Direction Sensor (CSL) measurements 'WS_ms' and 'WindDir'
PulseCount(WS_ms,1,1,1,1,0.098,0)
BrHalf(WindDir,1,mV2500,9,1,1,2500,True,0,60Hz,355,0)
If WindDir>=360 Then WindDir=0
'S2202/52203 Rain Gage (CSL) measurement 'Rain_mm'
PulseCount(Rain_mm,1,2,2,0,0.1,0)
'Call Data Tables and Store Data
CallTable(Table1)
CallTable(Table2)
NextScan
EndProg

Rabots glaciär logger program, version 2

'CR1000
'peter.jansson@2013–03–23

'Declare Variables and Units
Public BattV
Public PTemp_C
Public AirT05
Public RH05
Public AirT1
Public RH1
Public AirT2
Public RH2
Public SlrkWDN
Public SlrMJDN
Public SlrkWUP
Public SlrMJUP
Public WS_ms
Public WindDir
Public Rain_mm
Public SR50(2)
Alias SR50(1)=RawDist
Alias SR50(2)=SignalQuality
Public TempCorr
Public TempCorrDist
Units BattV=Volts
Units PTemp_C=Deg C
Units AirT05=Deg C
Units RH05=%
Units AirT1=Deg C
Units RH1=%
Units AirT2=Deg C
Units RH2=%
Units SlrkWDN=W/m^2
Units SlrMJDN=kJ/m^2
Units SlrkWUP=W/m^2
Units SlrMJUP=kJ/m^2
Units WS_ms=meters/second
Units WindDir=degrees
Units Rain_mm=mm
Units TempCorr=degrees C
Units TempCorrDist=mm

'Define Data Tables
DataTable(Table1,True,−1)
DataInterval(0,15,Min,10)
Sample(1,AirT05,FP2)
Sample(1,RH05,FP2)
Sample(1,AirT1,FP2)
Sample(1,RH1,FP2)
Sample(1,AirT2,FP2)
Sample(1,RH2,FP2)
Average(1,AirT05,FP2,False)
Average(1,AirT1,FP2,False)
Average(1,AirT2,FP2,False)
Average(1,SlrkWDN,FP2,False)
Totalize(1,SlrMJDN,IEEE4,False)
Average(1,SlrkWUP,FP2,False)
Totalize(1,SlrMJUP,IEEE4,False)
WindVector(1,WS_ms,WindDir,FP2,False,0,0,0)
FieldNames("WS_ms_WVT,WindDir_D1_WVT,WindDir_SD1_WVT")
Maximum(1,WS_ms,FP2,False,False)
Minimum(1,WS_ms,FP2,False,False)
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation

\[
\begin{align*}
\text{StdDev(1, WS, FP2, False)} \\
\text{Totalize(1, Rain, FP2, False)} \\
\text{Sample(1, RawDist, FP2)} \\
\text{Sample(1, SignalQuality, FP2)} \\
\text{Sample(1, TempCorrDist, FP2)}
\end{align*}
\]

EndTable

\[
\begin{align*}
\text{DataTable(Table2, True, -1)} \\
\text{DataInterval(0, 1440, Min, 10)} \\
\text{Minimum(1, BattV, FP2, False, False)}
\end{align*}
\]

EndTable

'SMain Program

BeginProg

'Main Scan
Scan(10, Sec, 1, 0)

'Default Datalogger Battery Voltage measurement 'BattV'
Battery(BattV)

'Default Wiring Panel Temperature measurement 'PTemp_C'
PanelTemp(PTemp_C, 0, 60Hz)

'HC2S3 Temperature & Relative Humidity Sensor measurements 'AirT05' and 'RH05'
VoltSe(AirT05, 1, mV2500, 1, 0, 60Hz, 0, 1, -40)

'HC2S3 Temperature & Relative Humidity Sensor measurements 'AirT1' and 'RH1'
VoltSe(AirT1, 1, mV2500, 3, 0, 60Hz, 0, 1, -40)

'HC2S3 Temperature & Relative Humidity Sensor measurements 'AirT2' and 'RH2'
VoltSe(AirT2, 1, mV2500, 5, 0, 60Hz, 0, 1, -40)

VoltSe(RH05, 1, mV2500, 2, 0, 60Hz, 0, 1, 0)

VoltSe(RH1, 1, mV2500, 4, 0, 60Hz, 0, 1, 0)

VoltSe(RH2, 1, mV2500, 6, 0, 60Hz, 0, 1, 0)

If RH05 > 100 AND RH05 < 103 Then RH05 = 100

If RH1 > 100 AND RH1 < 103 Then RH1 = 100

If RH2 > 100 AND RH2 < 103 Then RH2 = 100

'SR50 on SDI−12
If IfTime(0, 15, Min) Then
PortSet(9, 1) 'Turn on SW 12V
Delay(0, 3, Sec)
SD112Recorder(SR50(), 1, 0, "M1!", 1, 0)

TempCorrDist = RawDist + (SQR((TempCorr + 273.15)/273.15))

PortSet(9, 0) 'Turn off SW 12V
EndIf

'CS300 Pyranometer measurements 'SlrMJDN' and 'SlrkWDN'
VoltSe(SlrkWDN, 1, mV250, 7, 1, 0, _60Hz, 1, 0)

If SlrkWDN < 0 Then SlrkWDN = 0

SlrMJDN = SlrkWDN * 0.05

SlrkWDN = SlrkWDN * 5

'SCS300 Pyranometer measurements 'SlrMJUP' and 'SlrkWUP'
VoltSe(SlrkWUP, 1, mV250, 8, 1, 0, _60Hz, 1, 0)

If SlrkWUP < 0 Then SlrkWUP = 0

ShrMJUP = ShrMJUP + 0.05

ShrWUP = ShrWUP - ShrWUP

'05103 Wind Speed & Direction Sensor (CSL) measurements 'WS_ms' and 'WindDir'
PulseCount(WS_ms, 1, 1, 1, 0, 0.098, 0)

BrHalf(WindDir, 1, mV2500, 9, 1, 1, 2500, True, 0, _60Hz, 355, 0)

If WindDir >= 360 Then WindDir = 0

'SR50(2) Rain Gage (CSL) measurement 'Rain_mm'
PulseCount(Rain_mm, 1, 2, 2, 0, 0, 0.1, 0)

'Call Data Tables and Store Data
CallCheckTable(Table1)

Storglaciären logger program

'Sg−met−2013
'SG−met−2013.04−11

'Declare Variables and Units
Public BattV
Public PTemp_C
Public AirT05
Public RH05
Public AirT1
Public RH1
Public AirT2
Public RH2
Public SWin
Public SWout
Public LWin
Public LWout
Public LWInCo
Public LWOutCo
Public Albedo
Public NetSW
Public NetLW
Public Tot_net
Public Tot_in
Public Tot_out
Public NRtempC
Public NRtempK
Public e_AirT2
Public Sat_VP
Public WS_ms
Public WindDir
Public Rain_mm
Public SR50(2)
Alias SR50(1) = RawDist

75
Alias SR50(2)=SignalQuality
Public TempCorr
Public TempCorrDist

Units BattV=Volts
Units PTemp_C=Deg C
Units AirT05=Deg C
Units RH05=%
Units AirT1=Deg C
Units RH1=%
Units AirT2=Deg C
Units RH2=%
Units SWin=W/m^2
Units SWout=W/m^2
Units LWin=W/m^2
Units LWout=W/m^2
Units LWoutCo=W/m^2
Units Albedo=percent
Units NetSW=W/m^2
Units NetLW=W/m^2
Units Net_out=W/m^2
Units NRtempC=degrees C
Units NRtempK=K
'Units e_AirT2=kPA
'Units Sat_VP=percent
Units WS_ms=m/s
Units WindDir=degrees
Units Rain_mm=mm
Units TempCorr=degrees C
Units TempCorrDist=m

'Calibration constants
'Sensitivity uV/W/m^2; 1000/Sensitivity:
'SWin=15.53
'SWout=13.68
'LWin=11.07
'LWout=11.85
Const SWinCal=645.3915
Const SWoutCal=73.0985
Const LWinCal=90.3742
Const LWoutCal=84.3885

'Define Data Tables
DataTable(Table1 ,True,−1)
DataInterval (0,15,Min,10)
Sample(1 ,AirT05,FP2)
Sample(1 ,RH05,FP2)
Sample(1 ,AirT1,FP2)
Sample(1 ,RH1,FP2)
Sample(1 ,AirT2,FP2)
Sample(1 ,RH2,FP2)
Average(1 ,AirT05,FP2,FALSE)
Average(1 ,AirT1,FP2,FALSE)
Average(1 ,AirT2,FP2,FALSE)
Average(1 ,SWin,FP2,FALSE)
Totalize (1 ,SWin,IEEE4,FALSE)
Average(1 ,SWout,FP2,FALSE)
Totalize (1 ,SWout,IEEE4,FALSE)
Average(1 ,LWin,FP2,FALSE)
Totalize (1 ,LWin,IEEE4,FALSE)
Average(1 ,LWout,FP2,FALSE)
Totalize (1 ,LWout,IEEE4,FALSE)
Average(1 ,LWoutCo,FP2,FALSE)
Totalize (1 ,LWoutCo,IEEE4,FALSE)
Average(1 ,Albedo,FP2,FALSE)
Average(1 ,Tot_in,FP2,FALSE)
Average(1 ,Tot_out,FP2,FALSE)
Average(1 ,Tot_net,FP2,FALSE)
Average(1 ,LWinCo,FP2,FALSE)
Average(1 ,LWoutCo,FP2,FALSE)
WindVector (1 ,WS_ms,WindDir,FP2,FALSE,0,0,0)
FieldNames("WS_ms_S_WVT,WindDir_D1_WVT,WindDir_SD1_WVT")
Maximum(1 ,WS_ms,FP2,FALSE,FALSE)
Minimum(1 ,WS_ms,FP2,FALSE,FALSE)
StdDev(1 ,WS_ms,FP2,FALSE)
Totalize (1 ,Rain_mm,FP2,FALSE)
Sample(1 ,RawDist,FP2)
Sample(1 ,SignalQuality,FP2)
Sample(1 ,TempCorrDist,FP2)
EndTable

DataTable(Table2 ,True,−1)
DataInterval (0,1440,Min,10)
Minimum(1 ,BattV,FP2,FALSE,FALSE)
EndTable

'Main Program
BeginProg
'Main Scan
Scan(10,Sec,1.0)
'05103 Wind Speed & Direction Sensor (CSL) measurements 'WS_ms' and 'WindDir'
PulseCount(WS_ms,1,1,1,1,0.098,0)
BrHalf(WindDir,1,mV2500,7,1,1,2500,True,0,60Hz,355,0)
If WindDir>=360 Then WindDir=0
If IFTime(0,60,Sec) Then
' Default Datalogger Battery Voltage measurement 'BattV'
Battery(BattV)
' Default Wiring Panel Temperature measurement 'PTemp_C'

EndProg
A.2. Plotted 15 minute data for Rabots glaciär and Storglaciären
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation
The image contains several charts and graphs that appear to show various meteorological or environmental data over a period from May 9 to May 18, 2013. The charts include data for temperature (T (°C)), relative humidity (Rh (%)), solar irradiance (I (Wm^-2)), wind speed (sw (ms^-1)), wind direction (WD (°)), and possibly other variables such as elevation (z (mm)). Each chart seems to display measurements for different days, with data points marked on a timeline that spans from May 9 to May 18, 2013. The charts are likely used to monitor and analyze the variations in these environmental parameters over the specified period.
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation
Meteorological differences between Rabots glaciär and Storglaciere and its impact on ablation
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation
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Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation

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**T (°C)**

---

**Rh (%)**

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**I (Wm⁻²)**

---

**sw (m/s)**

---

**WD (°)**

---

**z (mm)**
Meteorological differences between Rabots glaciär and Storglaciären and its impact on ablation
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