Retrieval of snow characteristics by QuikSCAT backscatter measurements over North Slope, Alaska

Diplomarbeit in der Studienrichtung Meteorologie und Geophysik

zur Erlangung des akademischen Grades Magister der Naturwissenschaften

eingereicht an der Fakultät für Geo- und Atmosphärenwissenschaften
der Leopold-Franzens-Universität Innsbruck

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Innsbruck, Juli 2011
“It is a capital mistake to theorize without data.”

(Sherlock Holmes)
Abstract

The objective of this thesis is to determine the impact of soil state- and meteorological conditions on the backscatter behaviour at Ku-band radar frequencies in Nordic tundra regions and to explore the feasibility for retrieving snow water equivalent. The backscattered signal of QuikSCAT at the North Slope of Alaska was analysed for the years 2007-2009 with a focus on the beginning and end of both winters. The signal behaviour is interpreted in terms of soil state and meteorological parameters measured at automated weather stations which are operated by the University of Alaska Fairbanks. Complementary to the active microwave measurements, microwave emission data measured by AMSR-E are used to study the signal behaviour at the beginning of the snow accumulation period and to estimate the magnitude of snow events. Furthermore, the feasibility of using passive microwave grain size (obtained from the ESA DUE project GlobSnow) for supporting the inversion of QuikSCAT data in terms of snow water equivalent (SWE) is tested. In-situ SWE data, measured in late April, are used to study the sensitivity of the radar signal to SWE and to validate the results of the model inversion.

QuikSCAT backscatter shows a distinct response to freezing of the ground as well as to accumulation and melting of snow. At the beginning of winter, the backscatter coefficient decreased by about 2-5 dB caused by freezing of the ground. The decrease was more distinct in autumn 2007 than in autumn 2008. In 2008, an early snow event prevented the drop to lower backscatter values. The areas with a high percentage of wetlands show a pronounced difference compared to non-wetlands regarding freezing of the surface and early snow events. In winter 2007/2008, the QuikSCAT backscatter sensitivity to SWE was about 20 mmSWE/dB over most of the flat or hilly areas. The sensitivity to SWE in winter 2008/2009 was lower due to smaller grain size. The feasibility of using grain size retrieved from passive microwave (AMSR-E) data as input to SWE retrieval from QuikSCAT backscatter data was checked. Reasonable results were obtained for end-of-winter 2007/2008. However, SWE was significantly underestimated for end-of-winter 2008/2009 because the AMSR-E grain size was overestimated over the study region. As snow melt is important for hydrology, ecology and climatology at high latitudes, the main melt events of both winters were investigated more closely. QuikSCAT is able to
detect melting of snow accurately and is therefore applicable to estimate the melt onset and duration. Melt duration agreed with in-situ point measurements of snow ablation in spring 2007/2008. In spring 2008/2009 the melt duration differs by several days, most likely due to the different scales of the point measurements versus satellite data which are averaging over large areas.
# Contents

Abstract iii  

Contents v  

1 Introduction  
  1.1 Motivation 1  
  1.2 State of Research 2  
  1.3 Goals and Outline 3  

2 Theoretical background 5  
  2.1 Scatterometry 5  
  2.2 Physical properties of snow 6  
    2.2.1 Snow Properties 6  
    2.2.2 Electromagnetic properties of snow in the microwave region 7  
  2.3 Concept for SWE retrieval based on QuikSCAT backscatter data 9  

3 Study area description and data base 13  
  3.1 Study area 13  
  3.2 QuikSCAT 14  
  3.3 AMSR-E 18  
  3.4 Snow surveys 21  
  3.5 Automated weather stations 21  
  3.6 GlobSnow 24  

4 Sensitivity study for Ku-band backscatter of snow 25  
  4.1 Data basis 25  
    4.1.1 Snow survey data 25  
    4.1.2 QuikSCAT data 26  
  4.2 General analysis of QuikSCAT backscatter 31  
    4.2.1 Time series of backscatter 31  
    4.2.2 QuikSCAT backscatter and MW emission behaviour in winter 2007/2008 34
CONTENTS

4.2.3 QuikSCAT backscatter and MW emission behaviour in winter 2008/2009 .............................................. 38
4.3 QuikSCAT backscatter melt signature EOW 2007/2008 .......... 40
4.4 QuikSCAT backscatter melt signature EOW 2008/2009 .......... 42
4.5 QuikSCAT backscatter behaviour of soil freezing and early snow season in 2008 ............................................. 46
4.6 Backscatter sensitivity to SWE .................................................. 51
  4.6.1 Backscatter sensitivity obtained from forward model calculations 51
  4.6.2 Pre-snowfall backscatter signature .................................. 51
  4.6.3 Sensitivity in Winter 2007/2008 ........................................ 53
  4.6.4 Sensitivity in Winter 2008/2009 ........................................ 56
  4.6.5 Comparison of both years .............................................. 59
4.7 QuikSCAT backscatter response to wetlands during soil freezing ... 64
  4.7.1 Winter 2007/2008 ....................................................... 64
  4.7.2 Winter 2008/2009 ....................................................... 69

5 SWE retrieval in northern Alaska using QuikSCAT backscatter measurements .................................................. 75
  5.1 Data basis ............................................................................ 75
  5.2 SWE retrieval using passive MW grainsize as input .................. 77
    5.2.1 EOW 2007/2008 ......................................................... 77
    5.2.2 EOW 2008/2009 ......................................................... 79
    5.2.3 Comparison of computed SWE with field measurements for EOW 2007/2008 ................................................. 82
    5.2.4 Comparison of computed SWE with field measurements for EOW 2008/2009 ................................................. 85
  5.3 Grain size estimation using QuikSCAT \( \Delta \sigma^0 \) and in-situ data as input .............................................. 87
    5.3.1 Mean statistical relation between observed SWE and \( \Delta \sigma^0 \) .................. 87
    5.3.2 EOW 2007/2008 ......................................................... 87
    5.3.3 EOW 2008/2009 ......................................................... 89
    5.3.4 Differences in inversely calculated grain size and GlobSnow grain size .................................................. 89

6 Discussion and Conclusions .............................................. 93

Bibliography ......................................................................... 97

Acknowledgments ................................................................... 103

Curriculum Vitae .................................................................... 105
Chapter 1

Introduction

1.1 Motivation

The assessment of snow characteristics is a crucial point of information for many meteorological and hydrological applications. The snow cover (SC) is a very important parameter for the radiative budget and therefore also for the heat budget while the snow water equivalent (SWE) is fundamental for the estimation of run off at the end of winter (Berezovskaya et al. 2010). SC and SWE also serve as a climatic index and help to predict and estimate the impact of climate change. Temperature change, especially in high latitudes, is influenced by changes in the land surface albedo and snow plays a key role there (Lemke et al. 2007). Many global climate models predict the largest changes of air temperature at high northern latitudes (polar amplification). Therefore it is important to know the changes of SC and SWE in the polar regions.

As the polar regions are remote and therefore just a few weather stations exist, snow monitoring is mainly based on satellite data (Rees 2006, p. 1). For regular (daily) snow information, passive and active microwave sensors onboard of Earth observation satellites provide best capabilities as they can penetrate clouds. Additionally, they are able to penetrate dry snow and thus to provide estimates of the mass of snow lying on the Earth’s surface (Rott 1997). As the study area is located in gentle and undulating terrain at the North Slope of Alaska, low spatial resolution is not of major disadvantage, whereas a high temporal resolution is useful, QuikSCAT scatterometer was chosen for this thesis to learn about the information content of Ku-band backscatter data for snow monitoring.

A scatterometer has similar swath width as a passive microwave radiometer, but resolution enhancements work better for scatterometer (Long et al. 2001). Additionally, a active microwave satellite mission for calculating SWE (Cold Regions Hydrology High-Resolution Observatory, CoReH2O) was proposed to the European...
Space Agency as part of the Earth Explorer programme. Therefore active microwave measurements from space seems to be a future-orientated field of remote sensing of snow and ice. The work presented in this thesis is aimed of supporting the knowledge of the information content of radar backscatter for snow cover studies.

1.2 State of Research

In the past, passive microwave radiometry was mostly used to calculate SWE from space as this technique got its proof over decades. Extensive research has been done to develop algorithms to extract SWE from passive microwave data (Pulliainen and Hallikainen 2001; Tait 1998; Hallikainen and Jolma 1992). These algorithm usually use the difference in emissivity between two frequencies from the earth’s surface, normally 18 or 19 GHz and 37 GHz. SWE calculations by passive microwave radiometry are operationally since 1988. Parameters, which significantly influence the retrieval of SWE are in particular grain size and the land cover type. In particular areas with a high percentage of lakes, mountains and forests provide problems for application of this technique (Rees 2006, p. 150-151).

In recent years, the scientific community also investigated the feasibility of active microwave sensors to determine snow characteristics, including SWE. Scatterometers like QuikSCAT provide useful information to detect terrestrial snowmelt (Wang et al. 2008) or ice phenological events like timing of freeze-up and break-up and ice cover duration on lakes (Howell et al. 2009). Demanding on the frequencies, active microwave sensors are very sensitive to snow cover and have therefore a high potential for global snow cover monitoring (Nghiem and Tsai 2001). ERS-1 SAR was used to monitor freeze-thaw cycles of high latitude terrestrial ecosystems (Rignot and Way 1994) or other environmental conditions (Rignot et al. 1994).

Yueh et al. (2009) investigated the response of airborne Ku-Band radar backscattering to different vegetations types and to changes in the snowpack. They showed the ability of active microwave sensors to observe terrestrial SWE. To confirm, that vegetation and snow accumulation also have a influence to spaceborne active microwave measurements, a comparison of airborne data with QuikSCAT data was made in that study as well.

Former studies on the ability to extract SWE from spaceborne active microwave measurements were performed by Shi and Dozier (2000a,b). In this study, L-, C- and X-band multipolarization radar backscatter data from the Shuttle Imaging Radar-C mission were used to estimate SWE and other snow characteristics. Even though the SWE estimates were reasonable, more recent studies indicated that a combination of X- and Ku-band frequency are more appropriate to estimate SWE (Shi 2006).

In 2005, Rott et al. (2010) proposed the Cold Regions Hydrology High-resolution
1.3 Goals and Outline

The retrieval of snow characteristics in tundra areas by satellite-born active microwave measurements is promising. However, in some respects such as computing SWE or the sensitivity of the backscattered signal to different snow conditions the research so far has been limited. It is very important to understand more about the factors contributing to the backscattering behaviour in order to understand the temporal evolution of the backscattered signal. Therefore the QuikSCAT backscatter behaviour in relation to different snow, weather and ground conditions at the North Slope of Alaska are analyzed. As the QuikSCAT footprint is large and the backscattered signal is very sensitive to different aspect and slope angles, most of the selected study domain is rather flat or hilly. Additionally a semi-empirical radiative transfer forward model for extracting SWE based on single-channel backscattering measurements is tested. The results are compared to in-situ SWE measurements obtained from annual snow surveys. The goal of the investigations therefore were to investigate:

- In which way affect changing meteorolocial, ground and snow conditions the QuikSCAT backscatter behaviour in tundra areas of Alaska?

- Are there significant differences in the backscatter behaviour of wetlands compared to non-wetlands especially at the beginning of winter (ground freezing, first snowfalls)?

- Is it possible to obtain a good estimate of terrestrial SWE based on QuikSCAT backscatter measurements at the North Slope of Alaska?

The thesis is organized as follows. Chapter 2 provides theoretical background on radar backscattering of snow and ice and explains the way how SWE can be derived from QuikSCAT backscatter measurements. A description of the study domain and of the data base is given in chapter 3. In chapter 4 the results of the sensitivity studies are presented. The results of the SWE estimations applying...
the semi-empirical radiative transfer forward model are drawn in chapter 5. The discussion and the conclusions are provided in chapter 6.
Chapter 2

Theoretical background

2.1 Scatterometry

Scatterometers are radar sensors that usually are used to derive near-surface wind vectors over the ocean. However, scatterometer data have proven to be also very useful on a variety of other applications including snow and ice studies. Generally, scatterometers measure the backscattering cross section $\sigma$ of the surface area which is illuminated by the sensor antenna (Elachi and van Zyl 2006, p. 305-306). As scatterometers send microwaves towards the earth’s surface to measure $\sigma$, they are independent of solar illumination and are only weakly affected by atmospheric propagation conditions.

Figure 2.1 shows the principle geometry of a bistatic radar. $\sigma$ indicates the scattering cross section, $R$ is the range between target and sensor, $G$ is the gain of the antenna and $P$ is the transmitted or received power, respectively. On most remote sensing platforms, the transmitting and receiving antenna is identical. The radar transmits very short pulses of microwave radiation towards the ground surface. Each scatterer within the illuminated area contributes to the power received by the antenna. Equation 2.1 is referred to as radar equation and describes how the radar cross section for a point target is calculated.

$$\frac{P_r}{P_t} = \frac{\lambda^2 G_t G_r}{(4\pi)^3 R_t^2 R_r^2} \sigma$$  \hspace{1cm} (2.1)

In that equation, $\lambda$ refers to the transmitted wavelength of the radar. If the received power $P_r$ is returned from an area $A$ defined by the resolution of the radar, $\sigma$ is replaced by the radar cross section per unit area $\sigma^0$ (Henderson and Lewis 1998, p. 132). Equation 2.2 shows that $\sigma^0$ is a dimensionless variable referring to the backscatter of the normalized surface area.

$$\sigma = \sigma^0 A_0$$  \hspace{1cm} (2.2)
As values of $\sigma^0$ may cover a wide range in terms of power of the scattered signal, it is usually specified logarithmically in decibels. Equation 2.3 shows the conversion from linear $\sigma^0$ to logarithmic $\sigma^0$.

$$\sigma^0 [dB] = 10 \log_{10} \sigma^0$$ (2.3)

As scatterometers are generally used where high radiometric resolution is more important than spatial resolution, scatterometer data normally get averaged over a large area to obtain radiometric resolutions around 0.1 dB or better (Rees 2006, p. 69).

2.2 Physical properties of snow

2.2.1 Snow Properties

Pure snow is generally a mixture of ice crystals, liquid water and air. The following parameters are used to characterize the state of a snow pack:

- snow density $\rho_s$
- grain size
- SWE

Snow densities lie typically between 200 and 600 kg m$^{-3}$ for mature and around 100 kg m$^{-3}$ for freshly fallen snow. Wind and gravity compaction as well as thermal metamorphism increase snow density in time. In cold regions, freshly fallen snow as well as older snow have usually a lower density. To describe the internal structure of
2.2 Physical properties of snow

a snow pack, the grain size (or crystal size) is often used. The grain size is normally the mean radius of the ice crystals and is typically in a range of 0.1 - 3 mm. Within a snowpack grain size and snow density may show significant variations as a result of melt and refreeze cycles during a winter. In areas like the North Slope of Alaska, where the winter temperature is generally below zero degree Celsius, it is unlikely that the snow pack contains any liquid water during winter. At conditions, where liquid water is stored in the snow pack, the liquid water content plays a significant role in respect to the snow density, the microwave emission and backscattering properties.

\[
SWE = \frac{1}{\rho_w} \int_0^d \rho_s dz \quad (2.4)
\]

Equation 2.4 shows the relation of snow water equivalent SWE to the densities of water \(\rho_w\) and snow \(\rho_s\) and the snow depth \(d\). SWE specifies the total amount of water contained in a snow pack and is the essential parameter for the snow/hydrology processes. If all the ice in a snow pack is melted, SWE would be the mass of water contained per unit area drained by the snowpack. If the snow pack is homogenous in respect to snow density, this equation can be simplified to

\[
SWE = \frac{\rho_s}{\rho_w} d \quad (2.5)
\]


2.2.2 Electromagnetic properties of snow in the microwave region

To understand the main factors contributing to the backscatter signal, a short description of the interaction between microwaves and snow-covered ground is necessary. Figure 2.2 shows the propagation of microwaves through a snowpack for an active microwave sensor.

Equation 2.6 shows the main factors for the total backscatter \(\sigma'\) from snow covered ground received at the sensor. \(\sigma_{as}\) indicates the scattering at the snow / air interface, \(\sigma_v\) represents the direct snow volume scattering, \(\sigma_{gv}\) is the scattering term for interactions between the snow volume and ground and \(\sigma_g'\) represents the direct contribution of the backscatter from the ground surface (Figure 2.2). The magnitude of the different contributing factors depends on absorption and scattering properties and on the signal backscattered from the background medium (Rott et al. 2010).

\[
\sigma' = \sigma_{as} + \sigma_v + \sigma_{gv} + \sigma_g' \quad (2.6)
\]
For the retrieval of SWE, the direct contribution from the snow volume is the important parameter. Therefore, it is necessary that the microwaves are able to penetrate the snowpack. If we assume a homogeneous snow layer with snow depth \(d_s\), the transmissivity \(\tau_s\) can be expressed by equation 2.7

\[
\tau_s = \exp \left( -\frac{k_e d_s}{\cos \theta'} \right) \tag{2.7}
\]

where \(k_e\) is the volume extinction coefficient of snow and \(\theta'\) indicates the refraction angle in the snowpack. Absorption and scattering within a snowpack are responsible for the extinction according to equation 2.8

\[
k_e = k_a + k_s \tag{2.8}
\]

where \(k_a\) is the absorption coefficient and \(k_s\) is the scattering coefficient of the snow volume. Equation 2.9 specifies the absorption coefficient in terms of complex permittivity \(\epsilon = \epsilon' - i\epsilon''\) and the microwave wavelength in vacuum \(\lambda_0\).

\[
k_a = \frac{2\pi \epsilon''}{\lambda_0 \sqrt{\epsilon}} \tag{2.9}
\]

The real part \(\epsilon'\) of the permittivity of ice for the microwave region is nearly constant with a value of around 3.15 and thus, consequently, the real part of the permittivity of snow depends only on the snow density. Maetzler (1995b) specifies the empirical equation 2.10 for snow densities smaller than 450 kg/m\(^3\).
where the snow density $\rho_s$ is defined in $M g m^{-3}$. The imaginary part $\varepsilon''$ of the permittivity of snow determines the losses by absorption in the snow volume. It depends strongly on the liquid water content in the snowpack as shown in figure 2.3a. An important parameter to describe the magnitude of losses in a medium is the penetration depth which is also known as attenuation length. The penetration depth is the distance at which the radiation intensity inside the medium is reduced to $e^{-1}$ compared to the uninfluenced signal. Both absorption and scattering losses are taken into account (Rees 2006). Equation 2.11 shows, that the penetration depth is the reciprocal value of the extinction coefficient.

$$\delta_p = \frac{1}{k_e}$$  \hspace{1cm} (2.11)

As the absorption losses increase strongly with increasing liquid water content, the penetration depth decreases strongly with increasing liquid water content as shown in figure 2.3b for three different frequencies. Under wet snow conditions, extracting of SWE from a snowpack is therefore not possible. On the other hand, the strong decrease of backscatter due to increasing liquid water content enables the detection of snow melt. As for dry snow, the imaginary part of the permittivity is very low and thus, the absorption coefficient is also small. Typically values of penetration depth into dry snow are around $3 - 5 m$ at Ku-band and about $20 m$ for C-band. This also means, that propagation of microwaves in dry snow is dominated by scattering losses and therefore, deriving of snow depth or SWE by means of microwave scatterometry is possible only under dry snow conditions (Rott 1997).

### 2.3 Concept for SWE retrieval based on QuikSCAT backscatter data

To estimate SWE from backscattered VV-polarized QuikSCAT data, this thesis follows a formulation of Ulaby et al. (1984). In this concept, a semi-empirical radiative transfer forward model is used for describing backscatter of snow-covered ground $\sigma^t$ at a given frequency and polarization $pq$ (Nagler et al. 2011):

$$\sigma^t_{pq}(\theta_i) = \sigma^{as}_{pq}(\theta_i) + \sigma^V_{pq}(\theta_i) + T^2_{pq}(\theta_i) \left[ \sigma^G_{pq}(\theta_i) t^2(\theta_i) \right]$$  \hspace{1cm} (2.12)

$\theta_i$ is the incidence angle and $\theta_r$ is the angle of refraction. The other variables correspond to those used in equation 2.6. $T$ is the power transmission coefficient at the air - snow interface and $t$ represents the snowpack transmissivity. As the
Figure 2.3: Left figure: Real part $\epsilon'$ and imaginary part $\epsilon''$ of the permittivity of snow with a density of 500 kgm$^{-3}$ and a frequency of 5.3 GHz as a function of liquid water content (Rack 1995). Right figure: Penetration depth of microwaves in snow as a function of liquid-water content. (Ulaby et al. 1984)
contributions from snow-ground and ground-snow are small compared to the direct volume scattering contribution, they are included in the direct volume scattering term.

The transmissivity of the snowpack is expressed by

\[ t(\theta_t) = \exp \left( -\frac{k_e d_s}{\cos \theta_t} \right) \]  \hspace{1cm} (2.13)

where \( d_s \) is the snow depth. With SWE = \( \rho_s d_s \), the transmissivity can be expressed in terms of snow water equivalent.

\[ t(\theta_t) = \exp \left( -\frac{k_e \text{SWE}}{\rho_s \cos \theta_t} \right) \]  \hspace{1cm} (2.14)

Nagler et al. (2011) specifies equation 2.15 which expresses the volume scattering contribution where the ratio \( \omega = k_s/(k_a + k_s) = k_s/k_e \) is the scattering albedo.

\[ \sigma^V_{pg}(\theta_t) = T^2_{pg}(\theta_t) \left[ 0.75 \omega_{pg}(1 - t^2(\theta_t)) \cos(\theta_t) \right] \]  \hspace{1cm} (2.15)

Summarized, this results in equation 2.16 for the total backscatter of snow covered ground for a given frequency and polarization.

\[ \sigma^t(\theta_t) = \sigma^{as}(\theta_t) + T^2(\theta_t) \times \left[ 0.75 \omega \cos(\theta_t) \left( 1 - \exp \left( -\frac{2k_e \text{SWE}}{\rho_s \cos \theta_t} \right) \right) \right] \]  \hspace{1cm} (2.16)

Equation 2.16 can be inverted in order to extract SWE from \( \sigma^t \).

\[ \text{SWE} = -\frac{\rho_s \cos \theta_t}{2k_e} \ln \left[ \frac{\left( \frac{\sigma^t - \sigma^{as}}{T^2} \right) - 0.75 \omega \cos \theta_t}{\sigma^G - 0.75 \omega \cos \theta_t} \right] \]  \hspace{1cm} (2.17)

(Rott et al. 2010)
Chapter 3

Study area description and database

3.1 Study area

The study domain (figure 3.1) covers roughly an area of 250-by-250 km of Alaska’s Arctic Slope (AAS). It is bounded by the Brooks Range in the south and by the Arctic Ocean in the north. The elevation within the study area ranges from sea level to around 2600m. The northern parts are flat (referred to as Coastal Plain) while the middle part includes gentle hills and valleys (Foothill region). The southern part is characterized by mountain ridges of the Brooks Range (Mountain region) (figure 3.2). The coastal plain area in the north with many lakes is predestined to address the investigations of the backscatter behaviour to wetlands while the eastern part (east of -151°) of the study area (see figure 3.2) addresses the meteorological and SWE investigations as there is a yearly snow survey at the end of winter. Additionally, the eastern area has a very good network of automated weather stations.

Figure 3.1 shows the main land cover types in the study area. At the coastal plain area, sedge and herbaceous plants dominate the landscape while the foothill region is mainly covered by dwarf scrubs. The main part of the mountainous region is free of vegetation and is covered by rocks and barren land.

The Alaskan Arctic Coastal Plain is estimated being covered by wetlands by up to 83% (Hall et al. 1994). Hobbie (1980) estimated, that active freshwater lakes and ponds cover up to 40% of the surface. Most water bodies are shallow and have a maximum depth less than 2.0 m. There is a decreasing trend of the number of lakes towards the inland with a maximum within close to the coast (figure 3.1).

The primary recharge mechanism for the lakes is meltwater from the snowcover at the end of winter. Any additional recharge during the year is received from direct precipitation as hydraulic connections within the lake drainage network is
nonexistent as a result of high summer evaporation (Bowling et al. 2003).

Despite the large number of surface lakes, the annual precipitation is low. Figures 3.3a and 3.3b show the climate record of precipitation and temperature for two different sites located in the Coastal Plain and Foothills region over a year, respectively. The Coastal Plain has usually less annual precipitation than the Foothills, where most of the precipitation of the North Slope occur. Typically, annual precipitation amounts may range from $< 120 \text{mm}$ along the arctic coastline to $> 260 \text{mm}$ in the Foothills and mountainous areas with the maximum amounts of precipitation observed in the summer month (Cherry et al.). Precipitation falls normally in the form of snow from mid-September to mid-May. Only in the few summer month precipitation falls mainly in the form of rain, even if snow can fall throughout the year. Snow accounts for around 40% of the total annual precipitation (Kane et al. 2000). The amount and spatial pattern of the winter precipitation is discussed more thoroughly in chapter 4.1.1.

Mean annual surface air temperatures are well below freezing throughout the Alaskan Arctic and therefore, continuous permafrost exists throughout the North Slope of Alaska. Osterkamp et al. (1985) reported an average permafrost thickness of more than 600 m at the Coastal Plain area. The active layer (the layer of soil above the permafrost that thaws each summer and refreezes every winter) ranges from 25 - 100 cm.

The study domain is located in a very windy area with more than 60 days of average wind speed greater than $5 \text{ m s}^{-1}$ in the uplands and about 110 days in the coastal areas (Liston and Sturm 2002).

### 3.2 QuikSCAT

The SeaWinds scatterometer onboard the QSCAT satellite was launched in June 1999 and operated for over a decade until November 2009, when the bearings in the motor of the spinning antenna failed. The scatterometer provided normalized radar cross section ($\sigma^0$) at Ku-band (frequency = 13.4 GHz; wavelength = 2.2 cm) at both vertical and horizontal polarization using a conically scanning pencil-beam antenna (see figure 3.4a and 3.4b). The measurements were made of two constant incidence angles. The inner beam with horizontal polarization and an incidence angle of $46^\circ$ covered a swath of 1400 km while the outer beam with vertical polarization and an incidence angle of $54^\circ$ covered a swath of 1800 km. The scatterometer provided 90 % global coverage of $\sigma^0$ every two days with higher temporal sequence in high latitudes. As the polar regions were scanned several times each day, reconstruction of surface $\sigma^0$ at higher spatial resolution was possible.

In this study, QuikSCAT enhanced resolution products following the Scatterom-
Figure 3.1: Geographical map and land cover of study area (data from Multi-Resolution Land Characteristics (MRLC) Consortium (2001)), additionally the geographical locations of the COOP climate stations of Kuparuk and Umiat are marked (for climate record see figure 3.3)
Figure 3.2: Geographical map of intensive study area. Solid lines show major rivers; dashed lines represent approximate boundaries of the Coastal Plain, Foothills and Mountains regions. From Berezovskaya et al. (2007)
Figure 3.3: COOP climate record of temperature and precipitation between 1971 - 2000 for (a) Kuparuk station (70.32°, -149.58°, 20 m, Coastal Plain), (b) Umiat station (69.37°, -152.13°, 81 m, Foothills), data from Western Regional Climate Center (2010), for geographical location of the climate stations see figure 3.1
Study area description and data base

The figures 3.5a and 3.5b show the principal way how these enhanced resolution products are produced. The SIR algorithm is a true reconstruction algorithm providing resolution enhancement by combining multiple passes and extracting information from the side lobes of the measurement. The end products are provided in two different forms as egg or slice SIR images. For each footprint eighth individual \( \sigma^0 \) measurements are extracted. These measurements are called ‘slices’ and have a nominal pixel resolution of 2.225 km with an estimated effective resolution of around 5 km. The combined measurements of the eight ‘slices’ are called ‘egg’ measurements and have a nominal pixel resolution of 4.45 km with an estimated effective resolution of around 8 - 10 km.

In this study, egg-based SIR images were used as they contain less noise and are less sensitive to calibration errors (Wang et al. 2008; Long and Hicks 2005). All QuikSCAT data used for this study were ordered through the NASA Scatterometer Climate Record Pathfinder (SCP) website (http://www.scp.byu.edu/) which is a NASA sponsored project to develop scatterometer-based data time series to support climate studies of the Earth’s cryosphere and biosphere (Brigham Young University, Center for Remote Sensing 2010). Table 3.1 contains a brief technical summary of QuikSCAT.

<table>
<thead>
<tr>
<th><strong>Table 3.1:</strong> QuikSCAT technical information</th>
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<td><strong>Mission Description</strong></td>
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<td><strong>Instrument Description</strong></td>
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3.3 AMSR-E

Snow products derived from data of the Advanced Microwave Scanning Radiometer - Earth Observing System (AMSR-E) are used for comparative studies. AMSR-E
3.3 AMSR-E

Figure 3.4: QuikSCAT scatterometer, (a) Geometry of the QuikSCAT scatterometer, (b) Artist conception of QuikSCAT; both images from Long and Hicks (2005)

Figure 3.5: BYU products, (a) Overlap of QuikSCAT slices for a few consecutive pulses during a single antenna rotation and several antenna rotations at two different orbit locations, (b) Comparison of ‘egg’ and ‘slice’ SIR products; both images from Long and Hicks (2005)
onboard the Aqua satellite launched in 2002 (see figure 3.6) operates at six frequencies and two polarizations. The frequency band ranges from 6.9 GHz to 89 GHz resulting in spatial resolutions of about 5 km in the 89 GHz band and about 60 km in the 6.9 GHz band. The main observation targets of AMSR-E include precipitation, sea surface temperature, water vapour, wind speed, cloud liquid water, sea ice, snow cover and soil moisture. The antenna scans the earth’s surface at a constant incidence angle of 55 degrees and a swath of about 1450 km (EOS 2006). The spatial resolution of AMSR-E data is twice that of the Scanning Multichannel Microwave Radiometer (SMMR) and the Special Sensor Microwave/Imager (SSM/I) data. Furthermore, AMSR-E combines all the channels of SMMR and SMM/I in one sensor. Table 3.2 contains a brief technical summary for AMSR-E (National Snow and Ice Data Center 2011).

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<th>Table 3.2: AMSR-E technical information</th>
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<td>Bandwidth (MHz)</td>
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<td>Mean Spatial Resolution (km)</td>
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3.4 Snow surveys

In order to study the response of QuikSCAT backscatter to snow properties, spatially distributed in-situ snow cover data are required. The Water and Environmental Research Center (WERC) of the University of Alaska Fairbanks (UAF) perform annually in-situ SWE and snow depth measurements at the North Slope of Alaska since 2006. The data are available to the public. For this study, the SWE measurements of the years 2008 and 2009 were used. In the field campaigns, Gravimetric SWE were sampled and snow depth was measured (often referred to as double sampling) over 25-m by 25-m areas of dedicated sites. With this scheme, 50 snow depth measurements and five snow core measurements were made on a L-shaped transect with a 1-m sampling interval and 25 measurements on each leg of the L.

The measurements in both years took place in late April (EOW) until the first days of May. These measurements accounted for all precipitation which has fallen throughout the winter from October to April as there is usually no melting of snow before the end of April. Generally for both years, snow survey sites were chosen to represent snow characteristics over a wide range of vegetation and terrain conditions. SWE was measured at elevations ranging from 4 m to 1474 m.

Furthermore, the snow sites were grouped into Coastal Plain, Foothills and Mountains in order to determine regional average SWE (see figure 3.2). Generally, Coastal Plain sites are located below an elevation isoline of about 150 m. The discrimination whether snow is classified into Foothills or Mountains regions depends not only on elevation but also on the surrounding topography. Elevation alone is not representative in the uplands (Foothills and Mountains) because easy access to the snow survey sites is very important and therefore, most of the upland sites are located in valley bottoms.

In 2008, overall 113 snow survey sites were visited. 24 sites were located in the Mountains, 47 sites in the Foothills and 42 in the Coastal Plain.

In 2009, a overall of 143 snow survey sites were visited. 32 sites were located in the Mountains, 60 sites in the Foothills and 43 in the Coastal Plain. (Berezovskaya et al. 2008a,b, 2010)

3.5 Automated weather stations

The study domain in this thesis has an excellent network of automated weather stations compared to other arctic regions. They are operated by the University of Alaska Fairbanks. For this study, data from 17 automated weather stations within the study domain were used (see figure 3.7). The meteorological parameters of in-
terest were temperature, snow depth, surface temperature and soil moisture. 12 stations report all these meteorological parameters, four stations report only temperature and one station reports only snow depth. The stations are well distributed over the study area and cover therefore different landscapes and elevations.

Figure 3.8a shows an image of the snow depth sensor Campbell Scientific Sonic Ranger 50 A(SR50A). The SR50A is together with the SR50 the snow depth sensor of all snow-reporting weather stations within the study area. The only difference between the newer SR50A and the SR50 is the housing of the ultrasonic sensor and some additional output options. To calculate the snow depth, the sensor emits 50 kHz sound pulses and measures the elapsed time between the emission and return of the pulse. The SR50-series records measurements at one minute intervals and reports them averaged over a hour (Campbell Scientific 2011c).

Figure 3.8b shows a image of the air temperature / relative humidity sensor Vaisalla HMP45C which is implemented in the weather stations. A platinum resistance thermometer measures temperature over a range of -40°C to +60°C. The sensor reports hourly averages (Campbell Scientific 2011b). The non-aspirated sensor is installed in a height of two meters (WERC 2011b).

To measure the soil water content, the Campbell Scientific water content reflectometer CS616 soil-moisture probe is used. As the permittivity of water is significantly larger than for other soil constituents and the wave propagation velocity of the microwave signal of the reflectometer is dependent on the dielectric permittivity, the water content surrounding the soil-moisture probe can be measured (Campbell Scientific 2011a). The soil-moisture probes are installed horizontally in holes up to a depth of 40 cm and report hourly average values (WERC 2011b).

The soil surface temperature is measured with three individual thermistors of the 44033 YSI Series. They are installed at the top of the soil interface and report hourly average values (WERC 2011b).
3.5 Automated weather stations

Figure 3.7: Automated weather stations at the Alaskan North Slope used for this study printed on a color coded GTOPO30 elevation model; colorbar shows elevation in m; stations in black letters report temperature and snow depth, stations in red letters report only temperature and stations in blue letters report only snow depth.

Figure 3.8: Some instrumentation of the automated weather stations, (a) snow depth sensor Campbell Scientific SR50A (Ultrasonic distance sensor); image from Campbell Scientific (2011c), (b) Air Temperature/Relative Humidity sensor Vaisalla HMP45C; image from Campbell Scientific (2011b)
3.6 GlobSnow

The feasibility to use grain size retrieved from MW radiometer data as a priori information for SWE retrievals from active MW sensors is one of the questions investigated in the thesis. Essential information for extracting SWE out of the QuikSCAT backscatter is a good knowledge of the grain size, as the grain size is a key parameter for calculating the scattering albedo. For this purpose, grain size estimations retrieved by microwave radiometry which are provided by GlobSnow were analysed in this thesis. The GlobSnow project started in 2008 is funded by the European Space Agency (ESA) Data User Element (DUE). The main goals of GlobSnow are to provide long term datasets (15-30 years) for the Northern Hemisphere on Snow Extent (SE) and Snow Water Equivalent (SWE) for climate research purposes. Grain size estimations of snow are a by product of SWE processings (Finnish Meteorological Institute 2010).
Chapter 4

Sensitivity study for Ku-band backscatter of snow

This chapter provides a detailed analysis of the sensitivity of QuikSCAT Ku-band backscatter to meteorological, ground and snow conditions. Basis for the investigations are the SWE data from the snow surveys, the data from the automated weather stations and the backscatter data of the years 2007 to 2009. Additionally, the backscatter response especially at the beginning of winter is investigated in areas with a high proportion of lakes.

4.1 Data basis

4.1.1 Snow survey data

The SWE data of the snow surveys are fundamental for investigations regarding the sensitivity of backscatter to SWE and to other snow properties. Figure 4.1a and figure 4.2a depict the interpolated end of winter (EOW) SWE for the study domain for the years 2008 and 2009, respectively. The high densities of snow survey sites around Lat: 68.55° and Lon: -149.4° mark the Upper Kuparuk Watershed Site which is an intensive investigation site of the Kuparuk River Watershed Study (Kane et al. 2000). Important to note, SWE generally tends to vary strongly on a small spatial scale for both years.

In 2008, the mean snow density was highest in the Coastal Plain followed by the Foothills and the Mountains. The average snow depth was highest in the Foothills followed by the Coastal Plain and the Mountains. That resulted in the highest average SWE in the Coastal Plain closely followed by the Foothills. The Mountains had the lowest average SWE.

In 2009, the mean snow density was highest in the Coastal Plain followed by the Foothills and the Mountains. The average snow depth was highest in the Foothills
followed by the Coastal Plain and Mountains. That resulted in the highest SWE for the Foothills region followed by the Coastal Plain and the Mountains. Table 4.1 summarises the different snow parameters more closely for the individual areas of the study domain.

Generally, average end of winter SWE tends to be highest in the Foothills and Coastal Plain and lowest in the Mountains (Berezovskaya et al. 2008a,b, 2010). Compared to a 10 year average of the snow survey in 2008, SWE was overall beneath the average. While the Coastal Plain matched nearly exactly the long term amount of SWE (99.5 %), the Foothills (80 %) and in particular the Mountains (57 %) had less. 2009 was clearly above the long term average. The Coastal Plain and the Foothills (138 % and 137 %, respectively) were clearly above the 10-year average while the Mountains region (117 %) was just moderately above the average.

Some sites reported very high SWE values in both years, especially around Lat: 69° - 69.5° and Lon: 148.7° - 149.5°. Berezovskaya (2010) reported in a personal communication, that these sites were close to Happy Valley within the White Hills. Moreover, they are located in the drift zone and therefore, high SWE is measured each year. The high SWE variability within this small area resulted from wind-snow-terrain interactions. Because the representativeness of these high SWE values for the large scatterometer footprint is questionable, SWE values above 200 mm were not used for QuikSCAT data analysis. Figures 4.1b and 4.2b give the modified interpolated EOW SWE values for the study area. The modified data were used as in-situ data basis for the further investigations of the sensitivity of QuikSCAT backscatter to snow accumulation.

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<th>Table 4.1: Average snow depth, snow density and SWE within the study domain</th>
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<td>Coastal Plain</td>
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<td>Foothills</td>
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<td>Mountains</td>
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4.1.2 QuikSCAT data

In order to understand the behaviour of QuikSCAT backscatter to different soil- and meteorological conditions, it is necessary to analyse the backscatter values over
4.1 Data basis

Figure 4.1: Interpolated EOW snow water equivalent from snow surveys in mm at North Slope, Alaska, in 2008; (a) original data, (b) re-scaled data (data more than 200 mm SWE are removed to show more details in low accumulations areas); data from Berezovskaya et al. (2008a)

Figure 4.2: Interpolated EOW snow water equivalent from snow survey in mm at North Slope, Alaska, in 2009; (a) original data, (b) re-scaled data (data more than 200 mm SWE are removed to show more details in low accumulations areas); data from Berezovskaya et al. (2010)
Figure 4.3 and figure 4.4 show the spatial pattern of the VV-polarized QuikSCAT backscatter coefficient over the study area for six different time periods in winter 2007/2008 and 2008/2009. Roughly, the top panel in both figures assumed to be representative for the summer period with snow free conditions. More precisely, the left image of the top panel accounts for mid-summer conditions, the right image of the top panel corresponds to the end of summer. The middle panel and the left image of the bottom panel correspond to the winter period and the right image of the bottom panel addresses the melting period.

For all images, the high backscatter values in the south and south-east correspond to the mountainous regions. For both years, the backscatter in July is strongly dependent on vegetation, terrain type and open water within the footprint. Therefore, the approximate boundary between the Coastal Plain, the Foothills and the Mountains can be seen quite precisely (cf. figure 3.2). The Coastal Plain had the lowest backscatter values with about -15 dB which was probably caused by a high percentage of liquid water of thaw lakes. Additionally, the wetlands of the Coastal Plain have low biomass and probably also smooth surfaces (cf. figure 3.1). This also accounts for the relative low backscatter compared to the Foothills. The Foothills show higher backscatter with values of about -13 dB. It is suggested, that this mainly implies rougher surfaces and small scale topography. Furthermore, the temperature is slightly higher compared to the Coastal Plain (cf. figure 3.3) and combined with less open water areas the biomass in that region is higher (cf. figure 3.1). The fore slopes of the mountains are responsibly for the high backscatter in the mountainous regions with values above -10 dB.

Based on the July backscatter, a general feature for both years is the lower backscatter in September followed by a increase of backscatter over the winter with a maximum in both years in March and again a decrease until mid-May. This characteristic, especially with the increase over the winter month, indicate a general sensitivity of QuikSCAT backscatter to snow accumulation.
Figure 4.3: QuikSCAT VV backscatter coefficient in decibels for six different time periods in 2007/2008
Sensitivity study for Ku-band backscatter of snow

Figure 4.4: QuikSCAT VV backscatter coefficient in decibels for six different time periods in 2008/2009
4.2 General analysis of QuikSCAT backscatter

This section deals with the response of QuikSCAT backscatter to different meteorological conditions. The main investigations include freezing of the active layer of the permafrost, snow accumulation and melting of the snow pack. Additionally, short time melt and freeze processes of the snow surface were also investigated.

4.2.1 Time series of backscatter

In order to understand the investigations of QuikSCAT backscatter to different meteorological conditions properly, a few general explanations regarding the radar backscatter characteristic within a year are given. Figure 4.5 shows the temporal evolution of horizontal and vertical polarized QuikSCAT backscatter values at two different locations for two years. The North White Hills are located within the Coastal Plain area but close to the Foothills. Upper Kadleroshilik is located within the Foothills.

The HH-polarized backscatter is generally higher by 1-2 dB than the VV-polarized one. As Howell et al. (2009) indicated, this results mainly from the lower incidence angle of the HH polarization of QuikSCAT.

Another distinct feature is the sharp decrease in backscatter around mid May and end of April for the years 2007 and 2008, respectively. This decrease in backscatter is typical for the melting period. As Ulaby and Stiles (1981) indicated, liquid melt water in the snowpack is responsible for the decrease of backscatter. The large contrast of the permittivity of wet snow and air reduces volume scattering significantly. Moreover, liquid water is responsible for large losses due to absorption. This behavior of microwave backscatter was also evaluated in other studies like Wang et al. (2008) and Yueh et al. (2009).

The third striking feature is the decrease of backscatter at the beginning of the winter period around mid September. The decrease results from freezing of the active layer of the permafrost. As liquid water has a high permittivity, it is strongly responsible for the response of natural landscapes to microwaves. When freezing starts, the polar molecules of liquid water get bound in a crystalline lattice and as a result, the permittivity decreases significantly. The transition from liquid to frozen water results therefore in a decrease of backscatter of microwaves (Kimball et al. 2001; Rignot and Way 1994).

To relate the temporal evolution of QuikSCAT backscatter to snow accumulation, it is necessary to get an idea of the beginning of snow accumulation and the following temporal evolution of SWE. For this purpose, passive microwave data from AMSR-E were used. A slightly modified algorithm from Josberger et al. (1993) was used in order to calculate the brightness temperature gradient ratio $TB_{GR}$:
In this approach, $T_{B\,GR}$ is calculated from the 19 and 37 GHz vertically polarized brightness temperatures of AMSR-E. Josberger and Beauvillain (1989) measured snow extent and Josberger et al. (1989) showed the high correlation of the algorithm and SWE. In this thesis, the V-polarized AMSR-E data are used. Chang et al. (1987) and others showed, that the 37 GHz frequency channels from Scanning Multichannel Microwave Radiometer (SMMR) are more sensitive to snow parameters like SWE and snow extent than the 18 GHz channels which are comparable to the 37 GHz and 19 GHz channels of AMSR-E. Therefore, $T_{B\,GR}$ increases with increasing snow accumulation. As $T_{B\,GR}$ is also sensitive to comparatively thin snow cover, it is also a good indicator for the beginning of the snow accumulation period.

Additionally to $T_{B\,GR}$, sonic snow depth measurements were used to analyse the temporal evolution of snow accumulation. These measurements should be seen as a helpful tool in order to explain the temporal evolution of the QuikSCAT backscatter signal and other parameters dependent on snow accumulation. However, as comparisons with snow survey snow depth measurements showed, these measurements are often not representative for the surrounding terrain of the snow depth sensor. Due to frequent and strong winds at the North Slope, direct measurements of snowfalls can be even less reliable.

Figure 4.6a and figure 4.6b demonstrate comparisons of the sonic snow depth sensor with snow depth measurements from the snow survey at the surrounding area. In most cases, the snow depth measured by the sensor was underestimated compared to the average of the snow survey snow depths. As the differences between the observed surrounding snow depth values and the automated measured snow depth values vary from year to year, a correction of the automated measurements can not be applied. Additionally to the comparison of the snow depth measurements, the figures show the high variance of the snow course data. This shows the importance of using many measuring points per snow survey site (Berezovskaya et al. 2010).

Before the more precise description of the QuikSCAT backscatter behaviour during the winter period, it should be mentioned, that QuikSCAT VV backscatter is used for the rest of the study. There are two main reasons to justify that decision. As Yueh et al. (2009) indicated, in snowpacks with relative large vertical temperature gradients (e.g. Alaska) it is likely, that depth hoar can occur. As the facets of the snow grains in the depth hoar are oriented parallel to the H polarization, it may be possible that H polarization may have a stronger response to depth hoar than the V polarization. The second reason is the larger incidence angle of the V polarization which makes volume scattering from snow more effective.
4.2 General analysis of QuikSCAT backscatter

Figure 4.5: Temporal evolution of sigma nought in 2007/2008 at (a) North White Hills station, (b) Upper Kadleroshilik; Temporal evolution of sigma nought in 2008/2009 at (c) North White Hills station, (d) Upper Kadleroshilik
4.2.2 QuikSCAT backscatter and MW emission behaviour in winter 2007/2008

Figure 4.7a and figure 4.7b show the temporal evolution of the brightness temperature gradient ratio together with the air temperature and the QuikSCAT VV backscatter coefficient at the North White Hills station and Upper Kadleroshilik for a year during 2007/2008. To get a clear idea of the whole meteorological situation of the relevant time, figure 4.7c illustrates the snow depth measured at both stations. Additionally, the snow depth of Lower Kadleroshilik, located near the coast, is presented. To get a idea of the ground situation as well, figure 4.7d presents the temporal evolution of the surface temperature and soil moisture content over the year for the three stations mentioned before.

At the start of winter, around mid September, the already described decrease of sigma nought can be seen at both stations in figure 4.7. The reason for that feature was the decrease of air and surface temperature below zero degree Celsius which means that the ground began to freeze. The change of permittivity which resulted from the transition of liquid water to ice in combination with the measurement system led also to a apparent decrease of soil moisture content. The delay of the apparant decrease of around 25 days compared to the other parameters can be explained by the depth of 10 cm of the soil moisture sensors.

At the beginning of October, around 15 days after the freezing of the ground began, the brightness temperature gradient ratio at both illustrated stations increased significantly indicating snow covering the ground. Interestingly, QuikSCAT backscatter stayed low for about another 15 days at the North White Hills station and even 20 days at Upper Kadleroshilik. This indicates, that the amount of SWE at the
4.2 General analysis of QuikSCAT backscatter

**Figure 4.7:** Temporal evolution of QuikSCAT VV backscatter coefficient, brightness temperature gradient ratio and air temperature in winter 2007/2008 at (a) North White Hills station, (b) Upper Kadleroshilik, (c) Snow depth at three stations in winter 2007/2008, (d) Soil surface temperature and soil water content (-10 cm) at three stations in 2007/2008
Sensitivity study for Ku-band backscatter of snow

The snow depth measurements indicated just a thin snow cover by that time.

The first stronger snow events, especially in the North White Hills, were measured around mid October, when the ground was already frozen up to a depth more than 10 cm. The amount of fallen snow was sufficient to influence the radar signal, which began to increase from than on. A second sign of the larger snow amount on the surface, next to the snow depth measurements, is the stronger gradient of the brightness temperature gradient ratio. At Upper Kadleroshilik, the snow event was not as distinct as in the North White Hills. Therefore no increase of QuikSCAT backscatter was measured. A few days later the brightness temperature gradient ratio increased recognizably. Interestingly sigma nought also increased even though the snow depth measurements reported no distinct signal of fallen snow. A reason could be that both QuikSCAT and AMSR-E have a big footprint covering therefore not just the local measuring point at the weather station but the wide surroundings. If the weather station reports no snowfall, e.g. because of wind drift or local snow variability, there could nevertheless be fresh snow in the area around.

The next interesting time point was around mid January, when both QuikSCAT backscatter and the AMSR-E brightness temperature gradient ratio had a sharp and intensive decrease. By comparing the decrease with the air temperature, which is the mean of the maximum and minimum temperature measured at the weather station, it is evident, that a very short warm period occured. Within that warm period, surface melting may have taken place which was sufficient to decrease both parameters. Beside the decrease, the most striking feature is the strong increased backscatter at the North White Hills station following the warm period. As no snowfall appeared at that time, the strong increase must have other reasons. The most plausible suggestion is that snow crust was formed from melting and refreezing of the top layer of the snow. The large snow grains in that layer may have caused the significant increase of radar backscatter. This behaviour of radar backscatter was also evaluated in other papers like Yueh et al. (2009). A reason, why this increase from melting and refreezing can not be seen at Upper Kadleroshilik may be due to the higher level of backscatter which Upper Kadleroshilik in mid January already had compared to North White Hills. Additionally, this melt-freeze event was possibly less distinct at Upper Kadleroshilik. An evidence for this is the weaker QuikSCAT radar signal decrease during the melting days.

After that short warm period, the QuikSCAT backscatter stayed quite steady until the beginning of April when significant snowfall was reported from the snow depth sensors. Interestingly, just the radar backscatter at Upper Kadleroshilik showed a clear response to the reported snowfalls while the radar backscatter at the North White Hills maintained steady. As the brightness temperature gradient
ratio decreased at Upper Kadleroshilik significantly around that time, it can be assumed, that melt- and refreeze cycles led to the backscatter increase at Upper Kadleroshilik.

The next important event affecting the radar backscatter was at the end of April / beginning of May when a major snow event occurred which resulted in more than 10 cm of fresh snow at Upper Kadleroshilik and almost 5 cm at North White Hills. This snow event marks the heaviest in winter 2007/2008 according to the snow depth measurements. As the snowfall was heaviest at Upper Kadleroshilik, the radar backscatter increased there significantly in contrast to North White Hills, where just a slight increase was recognizable.

About mid May, for the first time since mid September, the air temperature stayed above zero degree Celsius for more than just a few days at both stations. Around this time was the main melt event within the study area. The radar backscatter reacted thereby with a large drop. This is due to liquid melt water within the snow volume. The whole snowpack was melted in less than 15 days. After the melt, around the beginning of June, the QuikSCAT backscatter increased to normal summer conditions. At that time, the soil surface temperatures were already above zero degree at all weather stations indicating the thawing of the ground.

The brightness temperature gradient ratio showed an interesting feature around the melt period. It dropped to the summer minimum in two steps. The first one was around 5 days before the main melting event began indicating thereby the high sensitivity of the brightness temperature gradient ratio to melt processes in the snow pack as the maximum day temperature was above zero degree at that time. The second drop, which was at the same time as the QuikSCAT backscatter drop, indicates the main melting days. A more thoroughly investigation of the QuikSCAT backscatter response during melting can be found in chapter 4.3.
4.2.3 QuikSCAT backscatter and MW emission behaviour in winter 2008/2009

In order to obtain additional information on QuikSCAT backscatter behaviour to meteorological events, the winter 2008/2009 was also investigated. This chapter provides focuses of meteorological events which were not investigated in previous chapters.

Figure 4.8a and figure 4.8b show the temporal evolution of the brightness temperature gradient ratio together with the air temperature and QuikSCAT VV backscatter at the North White Hills station and Upper Kadleroshilik for the winter 2008/2009. To get a clear idea of the meteorological situation for that winter, figure 4.8c illustrates the snow depth measured at both stations. The snow depth of Lower Kadleroshilik is also presented. To get a idea of the ground situation as well, figure 4.8d present the temporal evolution of the surface temperature and soil moisture content over the winter of the three stations mentioned before.

The winter 2008/2009 presented features different from to the previous one. Especially at the beginning of winter, the radar response was quite different compared to the winter 2007/2008. The strong decrease of backscatter which occurred mid-September 2007 was much less distinct in 2008. At Upper Kadleroshilik, sigma nought almost was steady throughout the whole pre-snowfall period while at North White Hills the decrease was much weaker compared to 2007. At both stations, the average pre-winter QuikSCAT backscatter minimum was weaker by about 2 dB compared to the year 2007.

When the first minor snowfalls arrived, a sharp increase of sigma nought was measured due to melt and refreeze cycles. This was quite different to the behaviour of the year 2007 where the backscatter increased more monotonically hand in hand with SWE. Another intriguing characteristic of the radar backscatter is the fairly steady course of radar backscatter at North White Hills and moreover the slight decrease of backscatter at Upper Kadleroshilik after that first strong increase of sigma nought. The very interesting feature especially at Upper Kadleroshilik was, that the backscatter decreased in connection with heavy snowfall between the beginning and mid-October.

This behaviour of backscatter is analysed in detail in chapter 4.5 for the time period around freezing of the ground in combination with the first snow events.

The next striking feature regarding the radar backscatterig series of the winter 2008/2009 is the melting period at the end of winter. Different from the previous year where melting lasted less than two weeks, the whole melt period took around two weeks longer. Moreover, melting was not associated with a temperature course rising steadily from below zero degree Celsius to above zero degree Celsius like in
Figure 4.8: Temporal evolution of QuikSCAT VV backscatter coefficient, brightness temperature gradient ratio and air temperature in winter 2008/2009 at (a) North White Hills station, (b) Upper Kadlerohilik, (c) Snow depth at three stations in winter 2008/2009, (d) Soil surface temperature and soil water content (-10 cm) at three stations in 2008/2009
the year 2008. In 2009, two main melt events can be identified.

The first, which took place around the last week of April, lasted around 5 days and was responsible for the melting most of the snow pack at the northern parts of the study domain (cf. snow depth in figure 4.8c). Additionally, it was responsible for a strong thawing of the ground which can be seen in figure 4.8d. After that first melt event, the temperature decreased around the beginning of May and additionally, snowfall occured in the second week of May.

The main melting began around mid-May which resulted in the typically ups and downs of the radar backscatter coefficient during melt and refreeze events as described in chapter 4.2.2 or in chapter 4.3.

Similar to the year 2008, the North Slope region was snow-free around the beginning of June which resulted in very high soil water content values at Lower Kadleroshilik. As the melt period of this year presents additional information on the QuikSCAT backscatter response to snow melt, chapter 4.4 provides a more detailed examination of the snow melt in EOW 2008/2009.

### 4.3 QuikSCAT backscatter melt signature EOW 2007/2008

In order to study the melting process thoroughly, figure 4.9 illustrates the melt process over the study area at the EOW 2007/2008. As the individual illustrations are averaged over three days, short time processes like the sharp peak from melt and refreeze processes can not be clearly seen. As result of averaging the gradual decrease and following increase of backscatter over time can be easily spotted. While the illustration before the 15th of May marks the state before the main melting with high backscatter of the snow pack, the main melt process started shortly afterwards which resulted in a drop of sigma nought between the 15th and 22nd of May throughout the study area (Coastal Plain and Foothills).

The temporal evolution of sigma nought for one point at the Coastal Plain (solid blue line) and one point at the Foothills (solid red line) is given in figure 4.10. Additionally the maximum and minimum air temperature of the Coastal Plain site Betty Pingo are shown.

The figure points out, that with air temperatures reaching above zero degree Celsius (solid black line, daily measured maximum temperature) the backscatter coefficient dropped at both areas (around 15th of May). Shortly afterwards, the air temperature decreased for a few days to values below zero degree Celsius. This led to refreezing of the snow surface at the Coastal Plain and consequently to the already discussed increase of backscatter at the northern areas as the grain size increased.
and the snow surface got more roughly. This increase of grain size led actually to a backscatter coefficient which was higher by about 1 dB compared to the one prior to the melt onset at some smaller areas. The maximum measured daily air temperature of the southern parts stayed above zero degree Celsius during these days which led to a steady ablation of SWE. When the daily measured maximum air temperature of Betty Pingo increased around the 22nd of May to temperature values of about 8 degree Celsius, the sigma nought characteristic of the Coastal Plain showes a large drop.

According to the spatial pattern of sigma nought (cf. figure 4.9), the complete melting lasted around 12 days. The melting started around the 15th of May at the Foothills and around the 22nd of May at the Coastal Plain. The discussed difference of the melt onset by about 7 days between the Foothills and the Coastal Plain region is apparent in this figure as well. According to the backscatter behaviour, the main melt lasted around 6 days in the Foothills and 3-4 days in the Coastal Plain.

In order to relate the backscatter behaviour to the SWE ablation, in-situ measured SWE ablation data of Berezovskaya et al. (2008a) are shown in figure 4.11. According to Berezovskaya et al. (2008a), the snowmelt onset in 2008 varied from May 15th at the Foothills (Imnavait Basin, Upper Kuparuk, Sagwon, Happy Valley) to May 21st - May 25th at the Coastal Plain (Franklin Bluffs, Betty Pingo and West Dock) with an average of 6 days for each region to complete the melt. Snowpacks across the entire study area melted away from May 15th to May 30th, 2008. The figure also illustrates the different melt behaviour of both areas. The SWE of the Foothills decreased steadily over a longer time range while the snowpack of the Coastal Plain was melted faster. The slightly higher air temperature at the southern area was responsible for the steadily ablation of SWE at the southern parts of the study domain.

These different time periods of the melt onset and melt durations matches the time periods retrieved by the backscatter coefficient.

A very interesting feature regarding the QuikSCAT backscatter response to snow melt is the large backscatter decrease of about 10 dB towards values lower than -22 dB in the Coastal Plain between the 24th and 26th of May. A reasonable explanation is that parts of the footprint were probably still covered by wet snow which decreased the backscatter. Additionally, the flat terrain at the Coastal Plain combined with the frozen ground below the surface layer may have allowed the melt water to stay there and built shallow thaw lakes. As thaw lakes generally cover big parts of the Coastal Plain and the radar backscatter reacts strongly to liquid water, the strong decrease is understandable. The soil water content of Lower Kadleroshilik as shown in figure 4.7d supports that explanation as well. The soil water content of Lower Kadleroshilik reached levels far above normal summer values
shortly after the melting period suggesting the building of shallow thaw lakes on top of the permafrost.

After the snow was completely melted around the 27th of May, the radar backscatter increased to summer values throughout the study area.

### 4.4 QuikSCAT backscatter melt signature EOW 2008/2009

The backscatter response to melting in EOW 2008/2009 was investigated as well. Figure 4.12 illustrates the melt process over the study area. As mentioned before, two main melt events were found in this year. The individual illustrations are
averaged over six days in order to get an idea on the whole melting period which took around two weeks longer compared to 2008. Due to the averaging, short time peaks of the backscatter coefficient can just be seen indirectly as a slight increase of the averaged backscatter coefficient. In order to show the detailed time sequence, the temporal evolution of \( \sigma_0 \) for one point at the Coastal Plain (solid blue line) and one point at the Foothills (solid red line) is shown in figure 4.13. Additionally, the maximum and minimum air temperature of North White Hills station are illustrated. In order to relate the backscatter behaviour to the SWE ablation, in-situ measured SWE ablation data of Berezovskaya et al. (2010) are shown in figure 4.14.

Figure 4.12 shows, that the first big snow melt event occurred at the end of April which resulted in a drop of the backscatter coefficient. The study area was affected by the melt. This drop can be clearly seen in figure 4.13 as well. The maximum temperature around that time reached values up to eight degree Celsius. Measurements of the ablation data started not before mid-May. Therefore, only the data of the snow depth measurements of the automated weather stations can be used (cf. figure 4.8c). This first melt event lasted about 4-5 days and was responsible for most of the ablation of snow pack at the northern parts of the study domain (see snow depth measurements).

Subsequently, the air temperature decreased well below zero degree Celsius which resulted in a strong increase of the backscatter coefficient due to the large
Figure 4.11: Net volumetric decrease in SWE. Snow ablation curves at the Foothills are shown as dashed lines and on the Coastal Plain as solid lines, EOW 2007/2008. Figure from Berezovskaya et al. (2008a)

grain sizes from the melt/freeze process. The biggest effects of the large grain sizes can be seen at the Coastal Plain, where the backscatter coefficient increased to values up to -8 dB which is high compared to the -12 dB before the melt occurred. Additionally, snowfall occurred in the second week of May which may have increased the backscatter as well.

On two minor occasions between the beginning of May and mid-May the maximum air temperature increased briefly above zero degree Celsius for a few days. This resulted on both occasions in a short-time drop of sigma nought (cf. figure 4.13). Especially the backscatter of the southern parts of the study domain were affected by that short-time melting as the northern air temperatures were slightly lower. This explains the big differences of sigma nought between the northern and southern parts around the second week of May (cf. 4.12).

At the end of the second week in May, the air temperature dropped once again below zero degree Celsius which resulted in the strong increase of backscatter over the study domain.

In mid-May, the main melt event eventually occurred throughout the whole study area. As shown in figure 4.13, the backscatter coefficient dropped thereby briefly
to values below -20 dB at the Coastal Plain which is quite similar to backscatter behaviour of the previous year and will therefore not be explained again. Due to the averaging, figure 4.12 shows a less distinct drop of backscatter. Regarding to the spatial pattern of the backscatter coefficient, the melt event was finished around the last week of May as no stronger backscatter change was observable. According to the backscatter behaviour, the melt of the whole snowpack throughout the study domain in 2009 lasted about four weeks if the snowmelt at the end of April counts as well. The main snow ablation took place in the second half of May and lasted around 7 days.

According to Berezovskaya et al. (2010), the 2009 snowpack was completely melted by May 26 at northern sites and around a week later at the southern Foothills. The survey stated, that the snowpack across the entire Kuparuk River basin was
It can be assumed, that the melt onset was detected by QuikSCAT quite accurately within a few days. Figure 4.14 starts later than the beginning of the snow ablation. The end of the melting period retrieved by QuikSCAT agrees with the in-situ data for the Coastal Plain area. In the northern Foothills, QuikSCAT retrieved snow-off dates are early by about a week.

A reason could be, that the minimum air temperature was below zero degree Celsius for the remaining main melt period and as the combined backscatter information of all daily passes of QuikSCAT were used, refreezing of parts of the snowpack at night may have prevented QuikSCAT to measure a clear signal if whether there was dry snow, wet snow or just wet soil. Moreover, Berezovskaya et al. (2010) presents ablation data of point measurements whereas the satellite averages over a large area.

### 4.5 QuikSCAT backscatter behaviour of soil freezing and early snow season in 2008

To illustrate the effect of soil freezing combined with subsequent snowfall on the backscatter at the beginning of winter 2008/2009, figure 4.15 shows the spatial pattern of QuikSCAT VV-backscatter coefficient for the study domain from the
4.5 QuikSCAT backscatter behaviour of soil freezing and early snow season in 2008

Figure 4.14: Net volumetric decrease in SWE. Snow ablation curves at the Foothills are shown as dashed lines and on the Coastal Plain as solid lines, EOW 2008/2009. Figure from Berezovskaya et al. (2010)

beginning of September to mid-October in 2008. All individual illustrations are averaged over five days.

Additionally, figure 4.16 shows the temporal evolution of the QuikSCAT backscatter coefficient over five month for individual footprints of the Coastal Plain (solid blue line) and the Foothills region (solid green line). Besides, the daily measured minimum and maximum temperature at Lower Kadleroshilik, which is situated in the northern part of the Coastal Plain, is also illustrated.

Figure 4.16 shows, that the backscatter coefficient at the Coastal Plain decreased quite strongly to values of around -19 dB after the 5th of September which is similar to the backscatter behaviour of the year 2007 before the snow accumulation began (cf. chapter 4.2.2). The reason for the decrease was similar to 2007 the freezing of the ground surface. However, at the Foothills no or just a slight decrease was recognizable which led to pre-snowfall backscatter coefficient values of about -15 dB.

It can be assumed, that the reason for the absence of a strong decrease at the Foothills was a slightly higher air temperature compared to the Coastal Plain in combination with a early snow event which occurred also around mid/end of
Sensitivity study for Ku-band backscatter of snow

Figure 4.15: Spatial pattern of QuikSCAT VV backscatter coefficient in decibels during the melt freeze cycle, averaged over five days, begin of winter 2008/2009

September. The short period between the mean air temperature reaching values below zero degree Celsius and snow covered the ground was too short to freeze the ground thoroughly (see figure 4.8). That resulted in relative high backscatter from the moist but unfrozen ground around mid-September. When the surface of the ground was eventually frozen between mid and end of September and the radar backscatter began to decrease in some parts of the Foothills (cf. figure 4.8a and 4.8b), the first snowfalls as shown in figure 4.8c appeared and stopped thereby the decrease of backscatter.

The spatial pattern of sigma nought in figure 4.15 reveals the different behaviour of the backscatter coefficient at the beginning of September as well. The individual illustrations indicate thereby the backscatter decrease at the Coastal Plain along
the steady backscatter course of the Foothills until mid-September.

Warmer temperatures characterized the following days (cf. figure 4.16) which resulted in thawing of the ground and therefore also in a sharp increase of the backscatter coefficient to values of about -15 dB at the Coastal Plain. It seems, that the thawing of the ground surface was so strong that the soil moisture sensors also responded to the warmer temperatures as the soil moisture content measured at 10 cm depth at Lower Kadleroshilik increased slightly at that time (cf. figure 4.8d).

Around the same time, the backscatter coefficient of the Foothills increased by about 2-3 dB as well. However, this increase possibly was not a result of ground thawing but mainly the result of an early snow event (cf. figure 4.8c) which was heaviest in the Foothills. In-situ measurements of the snow condition around that time are not available for this study but as the air temperature was around zero degree Celsius, it can be assumed that melt and refreeze processes led to a snow crust on top of the snowpack. This increase of sigma nought can be observed throughout the whole Foothills between the 20th of September and the beginning of October (cf. figure 4.15).

Additionally it can be assumed, that the snow cover of the Coastal Plain was quite shallow around that time as the backscatter dropped sharply to values of about -17 dB around the 20th of September indicating thereby, that the ground was frozen again. This could only happen, if the ground was just covered by a marginal snowpack. Otherwise, volume scattering of the snow would have increased the backscatter.

Around the beginning of October to mid-October, the backscatter coefficient at the Coastal Plain increased steadily to values close to -15 dB (see figure 4.15 and figure 4.16). The increase was mainly a result of snow accumulation as figure 4.8c demonstrates. The reason for the increase at the Coastal Plain was therefore similar to the previous year. On the contrary, the backscatter coefficient of the Foothills region decreased slightly (see figure 4.8b or figure 4.16) or remained steady (see figure 4.8a) despite of heavy snowfalls at this time. It is assumed, that the freshly fallen snow above the old snow crust from the melt-freeze cycle damped the strong backscattering signal from the big grains. Fresh snow has small grain size, which produces lower backscatter compared to snow with bigger grain size.

From that time on, snow covered the ground throughout the winter. In this year, the early snow cover dampened the heat exchange from soil to the atmosphere which resulted in the slow decrease of surface temperatures at all three stations shown in figure 4.8d. As a further result of the snow cover, the ground needed almost a month longer to freeze down to 10 cm compared to the previous year. Especially in Upper Kadleroshilik where the heaviest snowfalls occured the ground was unfrozen at 10 cm depth until mid-December.
Figure 4.16: QuikSCAT VV backscatter coefficient from Coastal Plain region (solid blue line) and Foothills region (solid green line), air temperature measured at Lower Kadleroshilik, end of 2008
4.6 Backscatter sensitivity to SWE

4.6.1 Backscatter sensitivity obtained from forward model calculations

In order to understand the behaviour of QuikSCAT backscatter sensitivity to SWE, forward model calculations of backscatter sensitivity to SWE have been performed. The model is specified in Nagler et al. (2011).

The backscatter sensitivity to SWE in tundra areas is mainly related to the pre-snowfall backscatter and to the effective grain size. Figure 4.17a and figure 4.17b show forward model calculations of backscatter sensitivity to SWE corresponding to pre-snowfall backscatter coefficients of -15 dB and -10 dB, respectively. Additionally, the calculations have been performed for three different grain size. As larger grains (0.6 mm) scatter more effectively at Ku-band the backscatter sensitivity is higher. The sensitivity drops gradually towards higher SWE values.

As SWE values at the North Slope are usually less than 250 mm, it has no strong effect on the further investigations of backscatter to SWE. Moreover, high pre-snowfall backscatter results in lower sensitivity values.

The forward calculations show a sensitivity of about 25 mmSWE/dB for a grain size radius of 0.6 mm and of about 55 mm/dB for a grain size radius of 0.4 mm for SWE $\leq$ 200mm and a pre-snowfall backscatter coefficient of -15 dB. The sensitivity drops to values of about 65 mmSWE/dB for a grain size radius of 0.6 mm and of about 85 mmSWE/dB for a grain size radius of 0.4 mm for SWE $\leq$ 200mm and a pre-snowfall backscatter coefficient of -10 dB.

4.6.2 Pre-snowfall backscatter signature

Generally, the sensitivity of QuikSCAT backscatter to SWE is calculated from the ratio of the QuikSCAT backscatter coefficient difference $\Delta \sigma^0$ in dB and the measured SWE in mm from the snow survey. $\Delta \sigma^0$ is the difference of the radar backscatter at the end of winter versus pre-snowfall backscatter.

Figure 4.18a and figure 4.18b illustrate the spatial pattern of sigma nought for both years over the study area before snow accumulation began. Both figures are averaged over 6 days. This ensures a stable sigma nought value which keeps the influence of short term fluctuations due to external factors to a minimum.

Most remarkably is the pronounced difference of the backscatter coefficient between level (Coastal Plain), undulating (Foothills) and mountainous terrain.

$\sigma^0$ values ranges thereby from around -18 to -16 dB in level regions up to around -9 dB in the mountains. This is mainly caused by the large footprint of the QuikSCAT sensor which covers a wide range of different slope and aspect an-
Figure 4.17: Backscatter sensitivity on SWE at 13.45 GHz (Ku-band), vv polarizations, 54° incidence angle, from forward calculations, for three different grain size and (a) pre-snowfall backscatter coefficient of -15 dB (b) pre-snowfall backscatter coefficient of -10 dB
4.6 Backscatter sensitivity to SWE

gles. Fore-slopes with high backscatter values in the footprint dominate thereby the mean backscatter as the angular dependence of backscatter is not linear. This causes problems in the mountains for the sensitivity of backscatter to snow accumulation which will be explained later.

The time-period for the pre-snowfall illustrations was chosen based on the characteristic of the brightness temperature gradient ratio. $TB_{GR}$ enables a good estimation of the beginning of snow accumulation over a wide area. This is also important for other studies like CoReH$_2$O which do not have high temporal resolutions and need therefore other ways to estimate the beginning of snow accumulation. Additionally, it is important where a lack of automated weather stations exist.

4.6.3 Sensitivity in Winter 2007/2008

The spatial pattern of $\Delta \sigma^0$ for EOW 2007/2008 are shown in figure 4.19. All individual illustrations were averaged over six days. As the snow survey data are gained around the last week of April, QuikSCAT data corresponding to this period were chosen. Besides, as short time melting with subsequent refreezing can influence the backscatter strongly, six different time-periods are presented.

As mentioned before and important to note, $\Delta \sigma^0$ not only is related to the accumulation of snow but also is a result of topographic effects and pre-snowfall conditions. This can be clearly seen in the mountainous regions, where $\Delta \sigma^0$ has values less than one dB. The highest $\Delta \sigma^0$ values can be found at the Coastal Plain and at the south-western parts of the Foothills with values of about 7 dB. These high values are observed for areas with low pre-snowfall values. The average of $\Delta \sigma^0$ near the centre of the Foothills was about 4 dB around mid-April and increased to values of about 5 dB at the beginning of May. This increase happened during the snow event at the end of April / beginning of May and was probably related to melt / freeze cycles of the surface layer.

The high increase of the backscatter coefficient around Lon: -147.2 and Lat: 69.6 was probably a result of melting and refreezing. Such high local increase in a rather short time could not be due to fresh snow. The same explanation can be accepted for parts of the south-western Foothills where sigma nought increased by about 1 to 1.5 dB within a few weeks.

To observe the sensitivity of the radar backscatter to SWE, the ratio of the snow survey data to $\Delta \sigma^0$ is calculated.

Figure 4.20 shows the sensitivity of the QuikSCAT backscatter coefficient to SWE in mmSWE/dB. To exclude very low sensitivity, sensitivity values less than 70 mmSWE/dB were removed and marked as no data (white color). The consequences can be identified in the mountainous regions which was removed almost completely.
Figure 4.18: Pre-snowfall QuikSCAT VV backscatter coefficient in dB over North Slope, Alaska, averaged over (a) 26 September 2007 - 01 October 2007 (b) 15 September 2008 - 20 September 2008
4.6 Backscatter sensitivity to SWE

Figure 4.19: Spatial pattern of QuikSCAT backscatter difference $\Delta \sigma^0$ in dB over North Slope, Alaska, averaged over six days, EOW 2007/2008 versus mean of 26/09/2007 - 01/10/2007 (pre-winter)
The high basic backscatter values caused from the fore-slopes make these regions almost completely insensitive to the QuikSCAT sensor regarding snow accumulation. The sensitivity in most parts of the Coastal Plain and Foothills areas is about 15 - 20 mmSWE/dB. At the area around the White Hills, sensitivity values of about 25 - 30 mmSWE/dB were calculated. This outcome is at least partly caused by the high snow survey SWE values in that region which may possibly be too high for the large area of the QuikSCAT footprint.

### 4.6.4 Sensitivity in Winter 2008/2009

In 2009, most of the study area had smaller $\Delta \sigma^0$ values (see figure 4.21) compared to 2008. Once again, all individual illustrations were averaged over six days. It can be assumed, that these smaller $\Delta \sigma^0$ values resulted mainly from smaller grain size in winter 2009. As SWE was higher in 2009, the temperature gradient was lower in the snowpack which supports the assumption, that grain size was smaller. Additionally, the higher backscatter coefficient at the beginning of the snow accumulation period (see figure 4.18b) contributed to the smaller $\Delta \sigma^0$ values.

The low values of $\Delta \sigma^0$ (around Lon: -149.5°, Lat: 69°) between the 18th April and 27th April at big parts of the Foothills resulted from short time melting. This melt event is also described in chapter 4.4. The following increase of the radar backscatter at these Foothills regions was mainly caused by refreezing of the snow pack.

The radar signal measured after the 18th of April was therefore affected by changes in grain size and does not represent the amount of snow on the surface accurately. Basically, $\Delta \sigma^0$ values were less than one dB in the mountains and up to six dB in some parts of the Coastal Plain. Interestingly, the spatial pattern of $\Delta \sigma^0$ at the EOW for the year 2009 has a higher variability compared to EOW 2008. This indicates, that in 2009 the SWE values or the grain size was more variable.

The spatial pattern of the sensitivity of QuikSCAT backscatter to SWE at the EOW for the year 2009 is shown in figure 4.22. Again, sensitivity values greater than 70 mmSWE/db were removed for scaling purposes. Comparisons of the spatial pattern of the sensitivity to SWE of 2009 with the year 2008 reveal, that the local variability of the sensitivity in 2009 is higher and differs spatially. Partly, this may be caused by the bigger amount of snow survey sites which results in higher variability of the interpolated SWE maps. Moreover, as mentioned before, the radar backscatter also shows a higher local variability which indicates the higher variability of SWE or grain size.

The melt event at the end of April influenced the sensitivity of the radar signal
4.6 Backscatter sensitivity to SWE

Figure 4.20: Spatial pattern of the sensitivity of QuikSCAT backscatter coefficients to SWE in mm SWE/dB, ratio of interpolated snow survey data versus QuikSCAT backscatter difference $\Delta \sigma^0$, EOW 2008, sensitivity values lower than 70 mmSWE/dB were removed in order to show more details in high sensitivity areas, white color marks no or removed data
Figure 4.21: Spatial pattern of QuikSCAT backscatter difference $\Delta\sigma^0$ in dB over North Slope, Alaska, averaged over six days, EOW 2008/2009 versus mean of 15/09/2008 - 20/09/2008 (pre-winter)
to SWE as well. This can be seen especially at the illustration of the time period between the 22nd of April and 27th of April. The values of Δσ₀ were such low at some areas of the south-western Foothills, that sensitivity values have been sorted out because of the allowed sensitivity minimum of 70 mmSWE/dB. Furthermore, the sensitivity increased strongly at the areas formerly affected by the melting as a result of the bigger grain size from refreezing.

As melt and refreeze cycles affected the radar backscatter after the 18th of April, these time periods are not considered for further investigations.

Roughly, the sensitivity values at the EOW 2008/2009 were about 25 - 30 mmSWE/dB at the Coastal Plain and decreased towards the centre of the Foothills to values of about 45 - 60 mmSWE/dB. The mountainous regions have been sorted out once again as the sensitivity is very low.

### 4.6.5 Comparison of both years

Histograms of the sensitivity values for both years are shown in figure 4.23. Figure 4.23a corresponds to EOW 2007/2008 while figure 4.23b corresponds to EOW 2008/2009.

Due to the melting within the second half of April in 2009, figure 4.23b shows the distribution of the sensitivity around two weeks earlier compared to figure 4.23a.

The mean values of the sensitivity was 19.5 mmSWE/dB for winter 2007/2008 and 35.4 mmSWE/dB for winter 2008/2009.

The big difference resulted mainly from two different influences regarding the radar backscatter and also the effective grain size.

Firstly, the radar backscatter was lower at the beginning of winter 2007/2008 compared to 2008/2009. As shown in chapter 4.6.1, lower pre-snowfall backscatter results in higher sensitivity. Secondly, it can be assumed, that the grain size was smaller in 2009. This would also lead to a lower sensitivity and is probably the main reason for the difference. Grain size will be discussed in chapter 5.3 more thoroughly.

Not only the mean values of the two years were different, but also the spatial distributions. The histogram of the sensitivity values at EOW 2007/2008 nearly matched a gamma distribution with the maximum of sensitivity of about 16 mmSWE/dB. The high standard deviation resulted mainly from the comparatively high sensitivity values near the boarder to the mountains.

The distribution for EOW 2008/2009 was different with the maximum of sensitivity of about 31 mmSWE/dB. Two significant levels of sensitivity can be identified with the borderrange between 30 mmSWE/dB to 34 mmSWE/dB. The left level with sensitivity values between 20 mmSWE/dB to 30 mmSWE/dB correspond
Figure 4.22: Spatial pattern of the sensitivity of QuikSCAT backscatter coefficients to SWE in mm SWE/dB, ratio of interpolated snow survey data versus QuikSCAT backscatter difference $\Delta\sigma^0$, for EOW 2009, sensitivity values lower than 70 mmSWE/dB were removed in order to show more details in high sensitivity areas, white color marks no or removed data.
mainly to the Coastal Plain. The right level with sensitivity values between 34 mmSWE/dB to 47 mmSWE/dB correspond mainly to the Foothills. Between the highest sensitivity values and the right level a sharp decrease can be seen. The highest sensitivity values (>47 mmSWE/dB) correspond on the one hand to the area between the Foothills and the mountains and on the other hand to the high values at the Foothills around Lon: -149.5 and Lat: 69.0.

In order to illustrate the differences of both years once again, figure 4.24a and figure 4.24b show a scatterplot of pre-snowfall backscatter versus backscatter sensitivity for both winter.

Figure 4.24a shows the pre-snowfall backscatter of September 2007 versus the retrieved sensitivity for EOW 2007/2008. The main pre-snowfall values (≤ -16 dB) correspond to the Foothills and the Coastal Plain. The main sensitivity values ranges from about 10 mmSWE/dB to values of about 23 mmSWE/dB. The figure shows, that with increasing pre-snowfall backscatter the sensitivity drops gradually as mentioned in chapter 4.6.1.

Figure 4.24b shows the pre-snowfall backscatter of September 2007 versus the retrieved sensitivity for EOW 2008/2009. Due to the assumed smaller grain size in winter 2008/2009 the sensitivity was lower compared to the previous year. Additionally, the pre-snowfall values of the Foothills and the Coastal Plain are not just concentrated on values of a few dB but cover a wide range of values (-20 dB - -14 dB, see also figure 4.18). The decrease of sensitivity with increasing pre-snowfall backscatter is evident in this figure as well.

It can be assumed, that the difference of the characteristics of both years is mainly caused by smaller grain size in winter 2008/2009. Additionally, the higher pre-snowfall backscatter may have played a role.
Figure 4.23: Histogram of sensitivity of QuikSCAT backscatter coefficients to SWE in mmSWE/dB for (a) EOW 2007/2008 (b) EOW 2008/2009

(a)

(b)

Mean: 19.5043 mmSWE/dB
Std: 10.9507 mmSWE/dB

Mean: 35.4237 mmSWE/dB
Std: 12.1764 mmSWE/dB
4.6 Backscatter sensitivity to SWE

Figure 4.24: Scatterplot of pre-snowfall backscatter versus backscatter sensitivity (a) mean backscatter of 26 September 2007 - 01 October 2007 versus backscatter sensitivity of 26 April 2008 - 01 May 2008 (b) mean backscatter of 15 September 2008 - 20 September 2008 versus backscatter sensitivity of 14 April 2009 - 19 April 2009
4.7 QuikSCAT backscatter response to wetlands during soil freezing

An interesting question regarding the sensitivity tests was also the response of QuikSCAT backscatter to wetlands within the footprint. This is especially important for possible further studies addressing the begin of ground freezing or SWE retrieval with active microwave sensors in areas with a high percentage of lakes.

If the time period of the pre-snowfall condition is calculated from the averaged backscatter (or e.g. AMRS-E data) over a large area (≫ lake-area), lakes within the area may be still unfrozen. As water can store energy longer than ground, the ground tends to freeze faster than lakes. This would lead to a pre-snowfall time period where the ground would be frozen and on the lakes would be still open water.

However, large unfrozen lakes within the footprint affect the backscatter. This could, for example, lead to a wrong estimation of SWE at the lake surroundings as the pre-snowfall backscatter will not represent frozen ground.

This sections discusses the sensitivity of QuikSCAT backscatter to a significant percentage of lake areas within the footprint.

4.7.1 Winter 2007/2008

The spatial pattern of the QuikSCAT VV backscatter signatures over the study domain can be seen in figure 4.25 for six different time periods at the beginning of winter 2007/2008. Every single illustration is averaged over five days.

The illustrations cover the backscatter conditions of pre-freezing, freezing and post-freezing of ground and lakes. The magenta box marks the wetlands while the black box marks a area with just marginal lake covering (further on referred to as non-wetlands).

No in-situ data of temperature and snow depth is available for the wetlands region for this thesis. However, it can be assumed, that the weather conditions were quite similar at both regions as both areas have a similar distance to the Beaufort Sea and are at about the same latitude. Therefore we can assume, that at both regions the temperature dropped below zero degree Celsius around mid-September (cf. air temperature figure 4.7a). The first snowfalls occured around the end of September / beginning of October.

The top panel of figure 4.25 shows the pre-freezing backscatter conditions. At this time, the two areas of interest had similar values. The Teshekpuk Lake around Lon: -153.5 Lat: 70.5 acts as a reference and should illustrate the time it took to freeze a large lake.
Further on, the ground began to freeze which resulted in a drop of radar backscatter at both regions (left illustration of mid panel). Interestingly, the drop of backscatter was stronger at the non-wetlands. A reason for that behaviour could be, that the wetland signal is strongly influenced by open water areas. Therefore, high winds around that time may have caused stronger backscatter from the lake surface compared to the frozen ground. The high backscatter at the Teshekpuk Lake and the Beaufort Sea supports this explanation. Additionally it can be suggested, that the ground had a higher liquid water content compared to the non-wetlands as well. This would also mean, that it took longer for the ground to freeze thoroughly which finally led to a more gradual backscatter response to freezing.

Subsequently, the first snowfalls occurred which increased the backscatter at both regions. As the Teshekpuk Lake still had very low backscattering values indicating open water, it can be assumed, that most of the lakes within the magenta box were unfrozen as well. Consequently, the snowfall had less influence to the backscatter over the wetlands as parts of the snow fell into open water. The fact, that the increase within the wetlands was strongest in the south-east (where less lakes are compared to the north-west) supports this assumption.

As for the bottom panel, completely frozen conditions (ground and lakes) can be assumed as the Teshekpuk Lake and even the Beaufort Sea began to freeze. Moderate snowfalls in the first week of October resulted in a further increase of backscatter at both regions. A reason for the small drop of backscatter after the 10th of October could be further snowfall which led to smaller effective grain size.

To get a better idea of the response of QuikSCAT backscatter to wetlands in the footprint, figure 4.26a shows the temporal evolution over four months for sigma nought for a single point within the wetlands (blue), the Teshekpuk Lake (red) and the non-wetlands (green).

Besides, figure 4.26b shows the difference of the point-measurements between the wetlands versus the non-wetlands. Therefore, negative values of the difference means lower radar backscatter at the wetlands. Once again, the backscatter of the Teshekpuk Lake acts as a reference for freezing of large water areas. The oscillation of radar backscatter from the Teshekpuk Lake indicate open water as the high amplitudes result mainly from strong winds.

This means, that the Teshekpuk Lake became frozen around the beginning of the second week of October. This supports the assumption made before, that the water areas of the wetlands were frozen around the beginning of October as the lakes within the magenta box are usually much shallower than the Teshekpuk Lake and freeze therefore faster.

Especially interesting to observe are the radar backscatter differences. It was
Figure 4.25: QuikSCAT VV backscatter coefficient in decibels, averaged over five days, the magenta box marks the wetlands; the black box marks the area without lakes, black circles mark comparatively big lakes, begin of winter period 2007/2008
significant in August with values of about -1.5 dB to -1.0 dB, got close to zero around the beginning of ground-freezing and was above zero before the snowfalls occurred. After the first snowfalls, the difference between both regions increased again to values of about -1.5 dB. Around the beginning of November, the difference decreased to values of about -0.5 dB.

As mentioned before, the difference in August was probably due to the high percentage of lakes.

The freezing of ground resulted in a backscatter coefficient difference of about zero as sigma nought dropped to values of about -19 dB at both regions.

The following increase of backscatter at the wetlands is one of the most interesting features. As mentioned before, the radar backscatter at the Teshekpuk Lake also increased significantly around that time. This indicates again the influence of strong winds to the backscatter at the lake-covered area. As strong winds only generated backscatter coefficient values of about -17 dB for the footprint where the Teshekpuk Lake is included, it can be assumed, that wind only has a strong influence on the backscatter behaviour of the wetlands when the backscatter from ground is quite low (e.g. frozen ground). Just than, the backscatter of the open water areas within the wetlands is important.

Subsequently, the first snowfalls occurred which led instantly to an increase of backscatter at the non-wetlands. As explained, the lakes were still unfrozen and therefore, the snowfall had less influence at the wetlands.

Further on, two snow events occurred during October. The first one was around mid-October and increased the backscatter at all three regions. It also decreased the difference of the backscatter values of the wetland region and the non-wetlands slightly. The second one occurred around the end of October and was responsible for the sharp increase of backscatter especially at the wetlands and the Teshekpuk Lake.

During November, the sigma nought characteristics of the three stations got comparable (moderate decrease of backscatter in the second week of November; increase of backscatter around the beginning of the last week of November) indicating thereby the strong decrease of the influence of the ground medium (water or ground).

In summary an important factor for the backscatter difference is the influence of water bodies. In the average, the difference was large in summer, decreased to values close to zero around freezing of the soil, increased again when the first snowfalls occurred and decreased afterwards step by step with each snow event. This indicates the influence of water areas within the footprint to the QuikSCAT backscatter. This is especially important if addressing freezing of ground and early snow events.
Figure 4.26: (a) Temporal evolution of QuikSCAT sigma nought over wetlands (blue), Teshekpuk Lake (red) and non-wetlands (green), (b) Temporal evolution of QuikSCAT sigma nought difference, wetlands minus non-wetlands; year 2007
4.7.2 Winter 2008/2009

As the soil/weather conditions at the beginning of winter 2008/2009 were different compared to the previous one, the wetland sensitivity study was also carried out during soil freezing in 2008.

Figure 4.27 shows the spatial pattern of sigma naught for six different time periods at the beginning of winter 2008/2009. The general set-up for the figure is similar to the one discussed before. The temperature dropped slightly below zero degree Celsius around mid-September before it started to increase again. It stayed continuously below zero degree Celsius after the 20th of September. The first snow events arrived around the last week of September (c.f. figure 4.8).

The left image of the top panel of figure 4.27 shows the conditions of the study domain during the first short freeze event. Similar to the previous year, the decrease in backscatter was less distinct in the wetlands. Once again, similar explanations to the ones discussed before can be made. As shown in figure 4.29a, the backscatter of the Teshekpuk Lake increased during the first freeze event indicating strong winds. This also led to a higher five days average of backscatter in the wetlands. Just on daily basis, the backscatter drop was quite similar in both regions (c.f. 4.29a).

Additionally, the higher liquid water content in the lake-covered area has probably dampened the backscatter decrease as well.

The following thawing of the ground around the 20th of September caused the higher backscatter values which are shown in the right illustration of the top panel of figure 4.27. The characteristic of sigma naught for the non-wetlands and the wetlands shown in figure 4.29a also demonstrate the sharp increase which the thawing caused. The increase of sigma naught was less distinct at the wetlands compared to the non-wetlands, as less ground is present and therefore able to thaw.

Further on, the temperature stayed below zero degree Celsius which resulted in the sharp drop of backscatter at both regions of interest.

The snowfall, which occurred in end of September, was responsible for the significant increase of the backscatter coefficient at the non-wetlands. It also caused the pronounced difference between both regions which can be seen in the left image of the middle panel of figure 4.27.

The WebCam pictures of two different lakes, which are stationed roughly between the two regions, are presented in figure 4.28. The WebCam pictures are online from spring 2008 on, therefore no WebCam picture for the previous year is available.

The left images (figure 4.28a and figure 4.28c) were taken before the lakes were frozen. The right ones (figure 4.28b and figure 4.28d) were taken after the freezing which occurred around the third of October. The top panel also shows the increase in snow cover at the lake surroundings between the first of October and the 09th of October. The following increase of radar backscatter after the beginning of October
Figure 4.27: QuikSCAT VV backscatter coefficient in decibels, averaged over five days, magenta box marks the wetlands; black box marks the non-wetlands, black circles mark comparatively big lakes, begin of winter period 2008/2009
4.7 QuikSCAT backscatter response to wetlands during soil freezing

Figure 4.28: WebCam pictures of two different lakes at the Coastal Plain, see timestamp for date, top pictures show lake conditions at Lon: 151.33 Lat: 70.23 (Lake L98 17), bottom pictures show lake conditions at Lon: -150.94 Lat: 70.33 (Lake L9312); (a, c) before freezing, (b, d) after freezing; WebCam pictures from WERC (2011a)

was therefore due to increasing SWE.

The interesting feature regarding the sensitivity of QuikSCAT backscatter to wetlands for 2008 is again the difference between the backscatter coefficient of the wetlands and the non-wetlands. It is presented in figure 4.29b for two footprint measurements. Similar to the previous year, the backscatter difference was large during summer and decreased towards the freezing period. When the first snowfalls arrived, the difference increased significantly to values larger than -2 dB due to the large unfrozen water areas at end of September.

Further on, the lakes began to freeze and therefore, the difference began to decrease as SWE accumulated on top of the lake ice. Additionally, the increase of ice thickness may have increased the backscatter over lakes as well.

In summary for both years it can be stated, that QuikSCAT backscatter is influenced by water bodies within the footprint. Differences between wetlands and
Figure 4.29: (a) Temporal evolution of QuikSCAT sigma nought over wetlands (blue), Teshekpuk Lake (red) and non-wetlands (green), (b) Temporal evolution of QuikSCAT sigma nought difference, wetlands minus non-wetlands; year 2008.
non-wetlands occur mostly after ground freezing (but lakes still unfrozen) in combination with early snow events. Significant differences happen to be also during summer. Additionally, the wind conditions are important to know over lake-covered areas when obtaining frozen ground backscatter conditions. Strong winds increase the backscatter noticeably when the surrounding ground is frozen.
Chapter 5

SWE retrieval in northern Alaska using QuikSCAT backscatter measurements

In this chapter the methodology for computing SWE from QuikSCAT backscatter measurements is described. A comparison of the resulting SWE values with field measurements is also given. Additionally, possible explanations for discrepancies between both SWE data sets are discussed.

5.1 Data basis

The basics of SWE retrieval using QuikSCAT backscatter measurements are discussed in chapter 2.3. SWE retrieval in this study is based on equation 2.17. Table 5.1 lists the parameters required for the equation. Additionally, basic explanations of the parameters are also included.

Furthermore, table 5.2 shows the values for the individual parameters used in this study. The individual parameters are described in chapter 2.3. The numerical values for the parameters used here are based on a snow density of $240 \text{kg/m}^3$, a snowpack mean temperature of $-12^\circ C$ and the QuikSCAT frequency of 13.4 GHz. The values of the individual parameters were calculated and obtained from Mag. Markus Heidinger, ENVEO IT GmbH, Innsbruck.
Table 5.1: Technical information for SWE retrieval using QuikSCAT backscatter measurements

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Name</th>
<th>Basis</th>
</tr>
</thead>
<tbody>
<tr>
<td>$k_s$</td>
<td>Volume scattering</td>
<td>Computed with Rayleigh scattering approximation</td>
</tr>
<tr>
<td>$T_s$</td>
<td>Mean snow pack temperature</td>
<td>For computing $k_a$, estimated from mean values of weather station data of soil surface- and air temperature</td>
</tr>
<tr>
<td>$\rho_s$</td>
<td>Mean snow pack density</td>
<td>Mean snow density for study region (from climate data), one averaged value for whole study domain</td>
</tr>
<tr>
<td>$\sigma^t$</td>
<td>Backscatter of snow covered ground</td>
<td>Obtained from EOW QuikSCAT backscatter measurements</td>
</tr>
<tr>
<td>$\sigma^G$</td>
<td>Backscatter of ground surface</td>
<td>Obtained from pre-snowfall backscatter measurements</td>
</tr>
<tr>
<td>$\omega$</td>
<td>scattering albedo</td>
<td>$\omega = k_s/k_e$</td>
</tr>
</tbody>
</table>

Table 5.2: Values of required parameters for SWE retrieval using QuikSCAT backscatter measurements

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Value</th>
<th>Additional declarations</th>
</tr>
</thead>
<tbody>
<tr>
<td>$k_s$</td>
<td>692993640.339522 * $r^3$</td>
<td>$r$=Snow grain radius in m</td>
</tr>
<tr>
<td>$k_a$</td>
<td>0.035478</td>
<td></td>
</tr>
<tr>
<td>$T^2$</td>
<td>0.998867</td>
<td></td>
</tr>
<tr>
<td>$\sigma^{as}$</td>
<td>0.002042=26.899452 dB</td>
<td></td>
</tr>
<tr>
<td>$\theta_t$</td>
<td>42.952°</td>
<td></td>
</tr>
<tr>
<td>$dmf$</td>
<td>0.952633</td>
<td>correction parameter for dense medium</td>
</tr>
<tr>
<td>$\omega_d$</td>
<td>$\omega \times dmf$</td>
<td>corrected $\omega$, $\omega$ replaced by $\omega_d$ in SWE-retrieval equation</td>
</tr>
</tbody>
</table>
5.2 SWE retrieval using passive MW grainsize as input

This section deals with the possibility of extracting SWE from QuikSCAT backscatter data. The needed grain size information for the study area was, amongst others, retrieved by microwave radiometry of the 19 and 37 GHz channels of AMSR-E in order to check the feasibility of using grain size from passive MW measurements for SWE retrieval. The synergy of passive microwave retrieved grain size information with active microwave backscatter is also checked in this chapter.

5.2.1 EOW 2007/2008

Correct grain size estimations are crucial for SWE retrieval based on QuikSCAT backscatter. \( k_s \), the volume scattering term, is dependent by \( r^3 \) with \( r \) being the radius of the effective grain size. Therefore, small changes in grain size may result in big changes of calculated SWE.

The grain size information from MW radiometric data used in this study was obtained from the Finnish Meteorological Institute (FMI) GlobSnow FTP server. The effective grain size estimations for EOW 2007/2008 are shown in figure 5.1. All individual illustrations are averaged over six days matching thereby the time-periods of the sensitivity study of QuikSCAT backscatter to SWE.

The spatial distribution of grain size within the study domain is quite homogeneous for any given date. On the contrary, the grain size estimations show significant temporal variability. The mean grain size diameter vary from 1.15 mm around mid-April to around 0.85 mm at the end of April/beginning of May.

The strong and sharp decrease of computed grain size in late April/early May is probably an artefact of the data set used for grain size retrieval which uses in-situ snow data of synop stations (anchor stations). The closest synop station to the North Slope is located in the south of the Brooks Range and may have reported different snow conditions compared to the ones of the North Slope. A further reason for the small grain size may be fresh snow from the snow event which took place in the last week of April (cf. figure 4.7c). The fresh snow may have added a new snow layer with a lower scattering albedo on top of the older snow. Transient melt events during this period which affected the mean brightness temperature may also play a role.

The retrieved SWE values at EOW 2007/2008 using grain size from passive MW are shown in figure 5.2.

Generally, the mountainous regions have the lowest SWE while the Foothills have the highest. The white space, which occurs in four illustrations around Lon:
SWE retrieval in northern Alaska using QuikSCAT backscatter measurements

Figure 5.1: Effective grain size estimation (diameter) in mm, GlobSnow data set, averaged over six days, EOW 2007/2008
5.2 SWE retrieval using passive MW grainsize as input

-147.5 and Lat: 68.8, corresponds to a area where the pre-snowfall backscatter was even higher than the EOW backscatter. This shows again the unfeasibility to calculate SWE out of QuikSCAT backscatter over complex terrain.

The most remarkable feature is the apparent strong increase of computed SWE in late April / early May over the study area. The apparent temporal changes in SWE are mainly determined by changes in effective grain size (cf. figure 5.1). The retrieved SWE values for the top- and middle panel of figure 5.2 vary between 90 - 130 mmSWE in the Foothills, around 90 mmSWE in the Coastal Plain and about 0 - 40 mmSWE in the mountains.

The calculated SWE values after the apparent grain size decrease (bottom panel) were around 170 mmSWE in the Coastal Plain, higher than 230 mmSWE in the Foothills and between 100 mmSWE and higher than 230 mm SWE in the mountains. The comparatively high SWE values in the mountainous region calculated for the 26th of April to first of May result from the short time increase of QuikSCAT backscatter (melt-freeze cycle) over this area (cf. 4.19) and the decrease of passive MW grain size.

5.2.2 EOW 2008/2009

The obtained effective grain size estimations for EOW 2008/2009 are presented in figure 5.3. Once again, the figure set up matches the figure set ups described before.

Similar to the previous year, the spatial distribution of grain size within the study domain is rather homogeneous for any given date. The temporal variability, like in the previous year, is also very pronounced in 2009. The mean grain size diameter is about 1.20 mm throughout the study domain until around the 23rd of April. Afterwards, a strong decrease in passive MW grain size can be observed.

The mean grain size decreased to values of about 0.85 mm in late April. Subsequently, the apparent mean grain size increased again to values of about 1.0 mm.

It can be assumed, that different snow conditions at the anchor stations led to the decrease of apparent grain size. Additionally, the increase of snow temperature and emitted radiance in the upper snow layer which is mainly effective at 36 GHz (due to smaller penetration depth than at 18 GHz) may have played a role.

The retrieved SWE values at EOW 2008/2009 for the study area are shown in figure 5.4. Once again, the figure match the set up of the SWE map described for EOW 2007/2008.

The SWE variability within the study domain is stronger compared to the previous year as the QuikSCAT backscatter had a stronger local variability. The strong increase of calculated SWE between the 22nd of April and first of May resulted mainly from the smaller computed grain size around that time-period. The follow-
Figure 5.2: SWE maps in mm derived of QuikSCAT backscatter data, averaged over six days, using passive MW grain size (GlobSnow) as input, EOW 2007/2008
5.2 SWE retrieval using passive MW grainsize as input

Figure 5.3: Effective grain size estimation (diameter) in mm, GlobSnow data set, averaged over six days, EOW 2008/2009
The calculated SWE values prior to the apparent grain size decrease vary between 20 - 60 mmSWE at the Coastal Plain, 40 - 100 mmSWE at the Foothills and 20 - 120 mmSWE at the Mountains. Subsequently, the calculated SWE increased to values between 100 - 180 mmSWE in the Coastal Plain, 140 mmSWE to more than 230 mmSWE at the Foothills and 80 to more than 230 mmSWE in the Mountains.

To analyse the effect of changing QuikSCAT backscatter to SWE retrieval, the time period between the 10th of April and 23rd of April (top panel and left illustration of middle panel in figure 5.4) is used as the grain size estimations were rather similar for that time period.

The area of interest is around Lon: -150° and Lat: 70° as the backscatter difference $\Delta \sigma^0$ (cf. figure 4.21) increased during this time by about 1 dB. Within the same time period, the calculated SWE values increased by about 20 mmSWE for that region. This results in a sensitivity of about 20 mmSWE/dB for that area within the Coastal Plain. This is comparable to the retrieved sensitivity values of about 25 mmSWE at the Coastal Plain which is discussed in chapter 4.6.4 (cf. figure 4.22).

### 5.2.3 Comparison of computed SWE with field measurements for EOW 2007/2008

This section deals with the comparison of computed SWE using QuikSCAT backscatter and passive MW grain size (GlobSnow) as input with in-situ measured SWE values obtained from the snow survey. The comparisons were done for both years.

The SWE retrieval algorithm using MW data should be applied before the melting period starts.

In EOW 2007/2008, the time period between the 18th of April and 23rd of April was chosen. The computed SWE using QuikSCAT backscatter and passive MW grain size as input for this time period can be seen in the left illustration of the middle panel of figure 5.2 in the previous section. Figure 5.5 shows the difference in mm of the computed SWE map minus the interpolated SWE map from the snow survey (cf. figure 4.1b). Additionally, figure 5.6a shows the histogram of the SWE difference for the same time period.

The mean difference is 2.78 mm which is rather low. The standard deviation with a value of about 34.39 mm is, on the contrast, rather big.

The spatial pattern of the difference of SWE reveal four zones of rather homogeneous differences. In the very north along the Coastline, QuikSCAT based SWE
5.2 SWE retrieval using passive MW grain size as input

![SWE maps in mm derived of QuikSCAT backscatter data, averaged over six days, using passive MW grain size (GlobSnow) as input, EOW 2008/2009](image)

**Figure 5.4:** SWE maps in mm derived of QuikSCAT backscatter data, averaged over six days, using passive MW grain size (GlobSnow) as input, EOW 2008/2009
Figure 5.5: SWE difference in mm, computed SWE using QuikSCAT backscatter and passive MW grain size as input minus interpolated in-situ measured SWE by the snow survey, EOW 2007/2008 values underestimate SWE significantly (up to 120 mm) due to the big QuikSCAT footprint. The Beauford Sea has a strong influence to the backscatter signal along the coastline. In the southern Coastal Plain, the computed SWE agrees well with the field measurements. The Foothills area shows a significant overestimation of computed SWE. The absence of spatial variability of passive MW grain size may be a reason for the overestimation. In the mountains, computed SWE underestimates SWE significantly due to the low sensitivity to snow accumulation.

Figure 5.6b shows the histogram of the difference of a computed SWE map with the in-situ SWE map after the apparent decrease of passive MW grain size in late April. As discussed in the previous section, computed SWE increased significantly after the apparent decrease of grain size. The decrease of grain size resulted in a mean difference of 136 mm and a standard deviation of 95 mm. This shows the high influence of grain size for computing of SWE.
5.2 SWE retrieval using passive MW grain size as input

Figure 5.6: Histogram of the difference of computed SWE and interpolated snow survey SWE in mm (a) prior to the apparent MW grain size decrease (b) after apparent MW grain size decrease; for EOW 2007/2008

5.2.4 Comparison of computed SWE with field measurements for EOW 2008/2009

In EOW 2008/2009, the time period between the 14th of April and 19th of April was chosen. The computed SWE results for this time period can be seen in the right illustration of the top panel of figure 5.4 in the previous section. Figure 5.7 shows the difference in mm of the computed SWE map minus the interpolated SWE map from the snow survey (cf. figure 4.2b). Additionally, figure 5.8 shows the histogram of the SWE difference for the same time period.

The mean difference is -68.61 mm and the standard deviation is 33.85 mm.

The spatial pattern of the difference shows the biggest differences of SWE at the Coastal Plain area with values up to 100 mmSWE. At the Foothills, computed SWE match the interpolated in-situ measurements of SWE well at some areas (red areas) while at others, the SWE difference is larger than 100 mmSWE (e.g. Lon -149.5, Lat: 69). The mountainous regions have the lowest differences of SWE compared to the other regions.

A reason for the strong underestimation of SWE is probably an overestimation of passive MW effective grain size. As shown before, computed SWE is strongly dependent on grain size. Besides, QuikSCAT backscatter captured the high SWE at the Coastal Plain in EOW 2009. Especially around Lon: -149.0 Lat: 70.0, where many snow survey sites were stationed, $\Delta \sigma^0$ had very high values (cf. figure 4.21). This supports the assumption, that errors in passive MW grain size were mainly responsible for the strong underestimation of SWE.
Figure 5.7: SWE difference in mm, computed SWE using QuikSCAT backscatter and passive MW grain size as input minus interpolated in-situ measurements from the snow survey, EOW 2008/2009

Figure 5.8: Histogram of the difference of computed SWE and interpolated snow survey SWE in mm, EOW 2008/2009
5.3 Grain size estimation using QuikSCAT $\Delta \sigma^0$ and in-situ data as input

In the previous sections, the grain size estimations were assumed to be mainly responsible for errors in computed SWE. This chapter shows results of an inversion of the semi-empirical radiative transfer forward model in order to estimate effective grain size from backscatter data. Additionally, it shows the differences of effective grain size obtained from the inversion with GlobSnow grain size.

5.3.1 Mean statistical relation between observed SWE and $\Delta \sigma^0$

In order to illustrate the variability of grain size, figure 5.9a and figure 5.9b show the difference between computed SWE from a mean statistical relation between observed SWE and $\Delta \sigma^0$ versus the SWE map interpolated from snow courses for the years 2008 and 2009, respectively. The computed SWE from the mean statistical relation (not shown) corresponds to a single value of effective grain size for the study area. This means, that the spatial variability of grain size is not taken into account which results in the differences shown in the figures.

In 2008, SWE is overestimated in the Foothills along the mountains which indicates, that in this area grain size was larger than the average. In parts of the Foothills, SWE is underestimated which indicates, that grain size was smaller than the average. In the mountains, SWE is underestimated as well. Due to the complex terrain, errors in computed SWE may not only be a result of errors in grain size but also because of errors in sensitivity as discussed before.

In 2009, SWE is mostly underestimated in the Coastal Plain and in parts of the Foothills (blue area) which indicates, that in this areas grain size was smaller than the average. At the White Hills and close to the mountains, SWE is overestimated (larger grain size).

As grain size has a significant variability, it is important to have a good information about grain size for computing SWE. The variability and magnitude of grain size will be discussed in the further chapters.

5.3.2 EOW 2007/2008

As input for the inversion of the semi-empirical radiative transfer forward model, QuikSCAT backscatter data from pre-snowfall conditions in 2007 (cf. figure 4.18a) and backscatter data from EOW 2007/2008 were used. Additionally, the interpolated in-situ map of SWE retrieved from the snow survey 2008 was used, assuming
Figure 5.9: Difference (in mm) between SWE map from mean statistical relation between observed SWE and $\Delta \sigma^0$ versus SWE map interpolated from snow courses, for (a) 18 - 23 April 2008 (b) 18 - 23 April 2009
5.3 Grain size estimation using QuikSCAT $\Delta \sigma^0$ and in-situ data as input

Figure 5.10: Map of effective grain size (diameter) in mm for 18-Apr-2008 - 23-Apr-2008, computed by inversion of the semi-empirical radiative transfer model, using QuikSCAT $\Delta \sigma^0$ and SWE from EOW snow survey data (interpolated map) as input, EOW 2007/2008 to represent correct SWE values for the study area. The spatial pattern of the retrieved effective grain size can be seen in figure 5.10. The retrieved grain size in the Coastal Plain is about 0.9 mm - 1.1 mm. In the Foothills, the estimated grain size is about 1.3 mm - 1.8 mm.

5.3.3 EOW 2008/2009

As input for the inversion of the semi-empirical radiative transfer forward model, QuikSCAT backscatter data from pre-snowfall conditions in 2008 (cf. figure 4.18b) and backscatter data from EOW 2008/2009 were used. Additionally, the interpolated in-situ map of SWE retrieved from the snow survey 2009 was used, assuming to represent correct SWE values for the study area. The spatial pattern of the retrieved effective grain size can be seen in figure 5.11. The retrieved grain size for most of the study domain is about 0.8 mm. In the Foothills along the mountains, the grain size is larger with values up to 1.3 mm.

5.3.4 Differences in inversely calculated grain size and GlobSnow grain size

The spatial pattern of the difference in inversely calculated grain size from QuikSCAT minus GlobSnow grain size for EOW 2007/2008 is shown in figure 5.12. As assumed in section 5.2.3 and section 5.3.1, GlobSnow grain size in large parts
SWE retrieval in northern Alaska using QuikSCAT backscatter measurements

Figure 5.11: Map of effective grain size (diameter) in mm for 18-Apr-2009 - 23-Apr-2009, Computed by inversion of the semi-empirical radiative transfer model, using QuikSCAT $\Delta \sigma^0$ and SWE from EOW snow survey data (interpolated map) as input, EOW 2007/2008

...of the Foothills is too low. The values of GlobSnow grain size estimations are about 0.2 - 0.5 mm lower than those obtained from QuikSCAT. Over most parts of the Coastal Plain, the grain size estimations match each other well which resulted in the accurate SWE calculations using passive MW grain size as input. As discussed in chapter 5.2.3, the high differences near the coast are due to the influence of the sea to the QuikSCAT backscatter.

The spatial pattern of the difference in inversely calculated grain size from QuikSCAT minus GlobSnow grain size for EOW 2008/2009 is shown in figure 5.13. As assumed in section 5.2.4 and section 5.3.1, errors in passive MW grain size was responsible for errors in computing of SWE. Passive MW grain size is overestimated over the whole study area. It is overestimated by about 0.4 mm - 0.5 mm over most parts of the Coastal Plain. In the south-western Foothills, passive MW grain size is overestimated by values up to 0.6 mm. In the Foothills along the Mountains and in the parts of the Foothills along the Coastal Plain, passive MW grain size is overestimated by about 0.2 mm - 0.3 mm.

It can be assumed, that GlobSnow grain size can not detect variability of grain size on a small spatial scale due to the large footprint of AMSR-E and due to the lack of anchor stations at the North Slope. This study points out that the grain size obtained from passive MW measurements is not applicable for input to SWE retrieval in this region.
5.3 Grain size estimation using QuikSCAT $\Delta \sigma^0$ and in-situ data as input

Figure 5.12: Spatial pattern of difference in inverted effective grain size (diameter) in mm, QuikSCAT backscatter based grain size minus GlobSnow grain size, EOW 2007/2008
Figure 5.13: Spatial pattern of difference in inverted effective grain size (diameter) in mm, QuikSCAT backscatter based grain size minus grain size of mean statistical relation, EOW 2008/2009
Chapter 6

Discussion and Conclusions

The QuikSCAT backscatter behaviour to different soil- and meteorological conditions in tundra areas was investigated over two years. Furthermore, a semi-empirical radiative transfer forward model was used to study the feasibility for estimating SWE from QuikSCAT backscatter data. Due to the need of abundant in-situ measurements for data interpretation and validation, the study domain at the North Slope of Alaska was selected. Even though mountains are covering part of the study area as well, a special focus regarding the investigations was laid on flat and hilly terrain.

In order to study the impact of changing soil- and weather conditions, data from 17 automated weather stations, operated by the University of Alaska, were analysed. These data sets include measurements of air temperature, snow depth, ground surface temperature and soil moisture content. In addition, AMSR-E data were used to detect the beginning of snow accumulation and to estimate the magnitude of snow events.

Generally, high sensitivity of QuikSCAT backscatter to different soil- and weather conditions was found. Due to freezing of the ground, the backscatter coefficient for comparatively dry ground at the beginning of winter 2007 decreased by 3-5 dB resulting in values of about -20 dB at the Coastal Plain and about -18 dB at the Foothills. In 2008, the decrease of the backscatter coefficient was less distinct as the first snowfall took place before the ground was thoroughly frozen. Therefore, the decrease of the radar backscatter coefficient was just about 2-4 dB resulting in values of about -19 dB at the Coastal Plain and about -15 dB at the Foothills.

The study area for analyzing QuikSCAT backscatter behaviour in the time around freezing included comparatively dry ground as well as Alaskan wetlands of the Coastal Plain. During summer, the QuikSCAT backscatter behaviour of the wetlands showed a similar temporal behaviour compared to the non-wetlands but was lower by about 1-1.5 dB due to a certain percentage of open water. When freezing of the ground began, the backscatter values of both regions got almost comparable due to the stronger decrease of backscatter of the non-wetlands. In
Discussion and Conclusions

both years, the first snowfalls occurred while the lakes were still unfrozen. The snowfalls resulted instantly in increase of backscatter in the non-wetlands. In the wetlands, the backscatter increased just marginally. Only when the water areas were eventually frozen and further snowfalls occurred, the backscatter values at both areas became comparable.

In order to study the sensitivity of QuikSCAT backscatter to SWE, data from annual field measurements at the end of winter for the years 2008 and 2009 were used. The in-situ SWE data were interpolated to retrieve SWE maps for both years. The interpolated SWE maps were used as ‘correct’ data for validation and were compared to the QuikSCAT backscatter coefficient difference $\Delta \sigma^0$. $\Delta \sigma^0$ is thereby the difference between the radar backscatter coefficient at the end of winter and the pre-snowfall backscatter coefficient in autumn. In EOW 2007/2008, sensitivity values of about 15 - 20 mmSWE/dB for most of the study domain were observed. Due to higher snow accumulation at the White Hills, suggesting the presence of smaller snow grains, these region had sensitivity values of about 25 - 30 mmSWE/dB. The observed values of sensitivity agree well with the sensitivity of 20 mmSWE/dB found by Yueh et al. (2009) for the North Park region in Colorado with snow cover on sparse grassland. In winter 2008/2009, an early snow event in the second half of September prevented the ground to freeze rapidly. Therefore, the decrease from pre-snowfall backscatter after snowfall was smaller than in the previous year. This and smaller grain size explain the contributed smaller $\Delta \sigma^0$ values compared to the year before, which resulted eventually in lower sensitivity values. For EOW 2008/2009, sensitivity values of about 25 - 30 mmSWE/dB for the Coastal Plain and about 45 - 60 mmSWE/dB for the Foothills were calculated.

As snow melt is important for hydrology, ecology and climatology at high latitudes, the main melt events of both winters were investigated more closely. In winter 2007/2008, snowpacks across the entire study area melted within 12 days between mid and end of May. Slightly lower temperatures at the northern parts of the study domain caused a later melt onset by about 7 days in the Coastal Plain compared to the Foothills. According to the backscatter behaviour, the final melting of the whole snowpack lasted around 6 days in the Foothills and 3-4 days in the Coastal Plain. It was found, that QuikSCAT backscatter retrieved time periods of melt onset and melt duration agree well with in-situ measured time periods. Additionally it was observed, that the melting caused a drop of the radar backscatter coefficient by about 10 dB to values below -22 dB at the Coastal Plain area. It is assumed, that this drop was a result of shallow thaw lakes from melted snow which developed at the Coastal Plain as the ground was still frozen below the surface layer. Additionally, wet snow still covering parts of the footprint may have decreased the backscatter as well. It was also observed, that short-term melt and refreeze processes of snow
resulted in QuikSCAT backscatter values which exceed those of dry winter snow measured prior to the melt onset. A similar result was also found by Wang et al. (2008) indicating significant impact of melt and refreeze events on the backscatter signal.

In EOW 2008/2009 two major melt events were observed. The first snow melt event occurred at the end of April and lasted about 4-5 days. It was responsible for major ablation of the snow pack in the northern parts of the study domain. The main melt event took place in mid-May and lasted around 7 days. According to the backscatter behaviour the melt of the whole snowpack throughout the study domain in 2009 lasted about four weeks. Similar to 2008, the melt onset was retrieved accurately by QuikSCAT. However, the end of the melting period retrieved by QuikSCAT was clearly evident only in the Coastal Plain area. Especially in the snow-prone northern Foothills, QuikSCAT retrieved snow-off dates show up earlier by about a week. It is assumed, that the meteorological conditions during the melt (minimum measured daily air temperature was below zero degree) is responsible for the early snow-off prediction by QuikSCAT. Under these conditions the average values over all daily passes of QuikSCAT which are used in this study include both wet and refrozen snow surfaces, impairing the clear discrimination between snow and bare soil.

Finally, an approach to retrieve SWE at tundra areas based on QuikSCAT backscatter measurements by inversion of a semi-empirical radiative transfer forward model was tested. The feasibility of using grain size estimations from the ESA funded GlobSnow project for model input was tested. The results were compared with the interpolated SWE maps from the snow surveys at EOW 2007/2008 and 2008/2009. In 2008, it was found that the mean model estimation of SWE agreed rather well with the in-situ SWE map. A mean difference over the whole study domain of 2.78 mmSWE was observed. However, the model estimations for the Foothills showed a significant overestimation of SWE which resulted in a standard deviation of 34.39 mmSWE. In 2009, large underestimation of SWE throughout the study domain was observed. A mean difference of -68.6 mmSWE combined with a standard deviation of 33.9 mmSWE between the model based SWE map and the interpolated in-situ SWE map was obtained. In 2009, the largest differences between the two SWE maps were found in the Coastal Plain. In this region, the model underestimated SWE by values larger than 100 mmSWE. This points out that the grain size obtained from passive MW measurements is not applicable for input to SWE retrieval in this region.

In order to estimate the magnitude of the error in passive MW grain size, the QuickSCAT measurements were inverted in terms of effective grain size using the semi-empirical radiative transfer forward model. QuikSCAT $\Delta \sigma^0$ and the in-situ
Discussion and Conclusions

EOE SWE map were therefore used as input to the model. In winter 2007/2008, GlobSnow grain size values were about 0.2 mm - 0.5 mm smaller than the values obtained by inversion of QuikSCAT data in the Foothills but agreed rather well at the Coastal Plain. In winter 2008/2009, GlobSnow grain size was overestimated over the whole study area with highest values up to 0.6 mm in some parts of the Foothills. This study points out that the use of passive MW grain size for input to SWE retrieval from QuikSCAT backscatter data is not suitable.

The investigations point out, that for SWE retrieval it is necessary to obtain representative information especially about the pre-winter soil conditions. These are crucial for sensitivity of backscatter to SWE and for the calculation of SWE by inverting backscatter measurements. This applied inversion model was found to be able to compute SWE reasonably well at flat or hilly tundra areas using QuikSCAT backscatter data as long as the grain size estimations are accurate. In order to further develop methods for retrieval of SWE by means of active microwave measurements, studies addressing the estimation of grain size are recommended. Moreover the studies confirm, that SWE retrieval results are reasonable as long as no strong melt event had occurred. Melting events have a large impact on the grain size also after re-freezing and therefore on the backscatter values as well.

Additionally it was observed, that wetlands have a different backscatter characteristic compared to non-wetlands regarding freezing of the surface or sensitivity to snowfall as long as the water areas of the wetlands are unfrozen. For further studies addressing the beginning of ground freezing by means of microwave backscatter measurements, it is therefore necessary to distinguish between wetlands and non-wetlands.

Besides it is suggested, that subsequent studies addressing snowmelt should use morning and afternoon passes of QuikSCAT even if Wang et al. (2008) found no evidence of significant diurnal variations of the QuikSCAT data during snow melt. The averaging over all passes during one day seems to be affected by night-time refreezing of the liquid water and is therefore not optimum for the estimation of melt duration.
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Acknowledgments

A great thank goes to several people without their contribution and support this thesis would not have been possible. First of all, I want to thank my advisor Prof. Helmut Rott for his support and assistance, the interesting discussions and all the time he had for me.

I am also thankful to members and former members of ENVEO IT GmbH, namely Mag. Markus Heidinger and Matze Reif for their support to this thesis. A great thanks goes to Mag. Florian Figwer who helped me with the first steps.

Further thanks goes to the Brigham Young University for providing all satellite data.

A special thanks goes to Sveta Stuefer and her team at the University of Alaska Fairbanks for the execution of the annual field campaigns and the helpful advices.

I appreciate the Finnish Meteorological Institute for providing the GlobSnow data sets.

I would also like to thank my friends at the university for the great time during my study. Further thanks goes to my flatmates and to my friends at home for the good time outside the university.

Finally, I am deeply thankful to my parents and my sisters for supporting me in every way all the time.
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