Measuring Surface Deformation Caused by Permafrost Thawing Using Radar Interferometry
Case Study: Zackenberg, NE Greenland

Master’s Thesis in Geomatics

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Title
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Abstract

Permafrost in high-arctic regions have been very much influenced by global warming. Many thousands of square kilometers of permafrost are under certain degree of thawing [46]. Site measurements have provided useful data for researchers. However, these ground-based measurements are usually site-specific with poor spatial coverage. Here we apply Multi-Temporal Interferometric Synthetic Aperture Radar (MT-InSAR) to measure the surface deformation over permafrost north-east Greenland during the past two periods 1995–1999 and 2006–2009 to infer permafrost behaviours. We have found a considerable rate of surface subsidence occurring with respect to a relatively stable area. Over the whole study area, during 1995–1999, we find a surface subsidence rate of 0.3–2.4 mm/yr and a seasonally varying displacements of 0.4–6.1 mm with subsidence occurring during the thawing season of each year. While in period of 2006–2009, we find the overall surface subsidence rate goes up to 0.8–2.7 mm/yr and the seasonally varying displacements increased to 2.3–7.4 mm within the thawing season. Comparing with the two periods, we find a general accelerating trend in subsidence rate of $4.5 \times 10^{-2}$ mm/yr$^2$, which indicates a acceleration in permafrost thawing rate. We also find a general increment in surface seasonal varying displacements of 1.9 mm, which indicates a thickening trend in the active layer.

In total, by applying MT-InSAR technique over the 5000 km$^2$ study area, we have found a permafrost area 506.1 km$^2$ of that is under thawing and the general magnitude of permafrost thawing rate is 17.0±8.4 cm/yr during 1995–1999. During 2006–2009, we have found 633.9 km$^2$ of permafrost area is thawing with a magnitude of thawing rate about 24.0 ± 12.0 cm/yr. We have also estimated the total volume loss of permafrost as $2.1 \times 10^8 \pm 1.1 \times 10^8$ m$^3$ and $3.0 \times 10^8 \pm 1.5 \times 10^8$ m$^3$ in each corresponding period.

Further, assuming the linear relationship between the thawed permafrost and the released methane fluxes observed during 19-Jun-1997 to 18-Aug-1997 at Zackenberg valley, we apply this relationship to the whole subsidence area. We estimate the total magnitude of methane fluxes released from the study area within this period, which is about $1081.7 \pm 398.5$ T.

This study provides new insights into the dynamics of permafrost systems and changes in permafrost conditions. It may further contribute to a better understanding of the permafrost-climate interaction. The results by InSAR are complementary to the traditional site measurements, thus this study also demonstrates the value using remote sensing technique alongside more traditional measurements.
This thesis is the last phase of my graduation from the Master of Science of Geomatics engineering. I started this two-year programme in September 2009. Generally speaking, I did quite well through my first year’s education. I have gained relative expertise in this field, including remote sensing technique, GIS management and GPS navigation. At that time, the Radar Interferometry technique didn’t draw much attention from me, since I didn’t have any background about that and I didn’t know what this technique can achieve. When the time came to choose my thesis topic, I followed my inclination to investigating on any kind of natural phenomena at a large scale, such as global warming, Tsunami, earthquake. This brought me to the Earth Observation department in the faculty of aerospace engineering. I still remember the first talk I had with Andy, he became my supervisor one week after. I told him about my interest and he introduced me to Radar Interferometry. I was quite intrigued when Andy showed me the latest results about observed deformation caused by the 2010 eruptions of Eyjafjallajökull in Iceland using this state-of-the-art technique. I was quite impressed by its capability of high spatial coverage with a few millimeters accuracy from 800 km altitude in space. After that talk, I decided to combine my interest with this fancy technique. This is how we came up with this topic to investigating on the permafrost thawing affected by global warming with Radar Interferometry.

To be honest, I would like to say it is quite a challenging topic for me. I started it without any prior knowledge of Radar Interferometry, nor in permafrost. Therefore, it really took me a very long time to understand how Radar Interferometry works and to gain knowledge about permafrost behaviors. When it comes to the interferometry processing phase, the problems followed one after another. At a time, I was really hesitated, frustrated and tired. However, I didn’t give it up half way. After nine months’ full-time hard working and struggling, I survived. More important, I have found some valuable information during this study.

It is at this point, I would like to take this opportunity to express my thanks and appreciation to all of you who have been supportive during my thesis work. First of all, I would like to thank my kind supervisor Andy Hooper for his great supervision and constructive comments. I also would
like to give my thanks to my graduation professor Ramon Hassen, who gave me very nice courses about Radar Interferometry and help me better understand the principle and such. I would like to thank the whole Radar Group, everyone is so kind and willing to help. I would like to thank the Zackenberg research team and Mikkel Peter, who provided me with the specific information in Zackenberg Valley and their site measurements data. I would like to thank my classmate Penelope Rammos and all other fellow students, who help me check my english grammar in writing. I would very much like to give my special thanks to Miguel C.Cuenca for all his first-hand aid when I really got stuck, for all his patience in answering my questions, for all his inspirations when I felt down and for all his valuable feedbacks. It would have been more difficult without his help along the way.

Last but not least, I would like to thank my parents and my whole family members. I am extremely grateful to my great father and mother. It is because of your full supports from behind that I do not feel alone. It is because of your full supports that I picked up my courage when I got lost. It is because of your full supports that made my study abroad possible. I thank my mother for giving me life and for raising me. I thank my father for teaching me skills and training me to become a man. I am really proud to be your son through my entire life.

Thank you everyone, thank you the Netherlands!

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Chapter 1

INTRODUCTION

This chapter describes the motivation of this study. It starts with some background information of this thesis and then defines the main objectives of this research. It continues with research methods and expected deliverables. A layout of this thesis is given at the end of this chapter.

1.1 Background

Permafrost has a large presence in the northern hemisphere. According to the International Permafrost Association (IPA), one quarter of the land area (about 23 million km$^2$) in the northern hemisphere is underlain by permafrost \[63\]. Permafrost is a thermal condition of the ground in which the temperature remains below 0$^\circ$C for at least two consecutive years \[52,72\]. It may comprise soil, sediment, ground-ice and organic materials. Therefore, permafrost environment is such a huge reservoir for Soil Organic Carbon (SOC). Researchers have estimated that permafrost soil in boreal and arctic ecosystem stores almost twice as much carbon as is currently present in the atmosphere \[62,79\].

Global warming is one of the most important issues that humans are confronted with today. Particularly, the magnitude of warming in the high-arctic is likely to be two to three times larger than in temperate regions \[43\]. In the polar region, there has been a rise in temperature at high latitudes of 0.3 – 0.6$^\circ$C during the last century \[46\]. Permafrost might be affected seriously by global warming. It is subjective to a thawing process, and this change in permafrost triggers greenhouse effect by letting methane escape into the atmosphere. Evidences of current change in permafrost are collected from different site measurements. Whalen \[74\] found a mean methane flux of 52 mg · m$^{-2}$ · d$^{-1}$ from arctic tundra in Alaska. In the period 1993 to 1995, Nakano \[53\] investigated a Siberian permafrost area, the waterlogged wetlands site at Tiksi, which displayed an averaged methane flux rate of 46.3 mg · m$^{-2}$ · d$^{-1}$. In the year 1997, Christensen \[17\] reported a methane
flux rate of 45.6 mg·m$^{-2}·d^{-1}$ in Zackenberg valley near northeast Greenland.

Regionally, a few degrees rise in temperature and the consequent melting of permafrost would influence the geomorphic, hydrologic and biological processes [54,73]. The increased air temperature and ground temperature might mobilize the carbon currently trapped in permafrost soil and trigger a higher emission in greenhouse gas. The released gas from thawing permafrost has become an additional source of greenhouse gases and poses a positive feedback towards global warming [63,64,79].

1.2 Research question and motivation

It is necessary to understand permafrost behaviors and its current changes. The reason is simple but quite strong. Firstly, permafrost thawing releases trapped carbon in the form of greenhouse gas and further enhancing global warming [45,79]. Secondly, the thawing permafrost causes surface deformation. It may alter the landscape, topography and cause damage to the infrastructure [2,19,68]. Thirdly, permafrost thawing disturbs the equilibrium of the regional ecosystem [73].

Permafrost-climate interaction is poorly understood due to the inaccessibility of most permafrost areas as well as the limited spatial coverage of site measurements. One aspect of interest is what degree of deformation may be expected if historically frozen permafrost soils start to thaw. Another interesting aspect is the magnitude of methane flux released from the vast thawing permafrost region.

With these common interests, the research question in this study is defined as: what is the on-going changes inside permafrost and what is the magnitude of these changes.

The objective of this study is to investigate the long-term permafrost changes by measuring the corresponding surface deformation and then to estimate the magnitude of released methane flux due to the amount of permafrost changes. This research provides the opportunity to give an inside view on the permafrost behaviors and the current changes. Further, it may contribute to the better understanding of permafrost-climate interaction.

Radar Interferometry, also known as Interferometry Synthetic Aperture Radar (InSAR), is a powerful tool in geodetic applications [33]. It has been widely used in ground surface deformation monitoring because of its ability to remotely sense centimeter to millimeter surface deformation over a large area with a spatial resolution in the order of tens of meters [14,33]. In this study, InSAR is the chief geodetic tool applied to measure the permafrost-related surface deformation over the study area, where it cannot be easily accessed with field work.
1.3 Research methods

In this study, we divided the whole work-flow into three parts. Different strategies are applied for each of these. The first part is concerned with the InSAR surface deformation measurements. We apply the advanced MT-InSAR method to obtain the time-series surface deformation, a method called the Small Baseline approach \cite{37}. The second part is about deformation modeling. The aim is to model the observed InSAR measurements with a deformation model. Two kinds of deformation signals can be obtained after modeling: one is the long-term surface deformation rate, another is the surface seasonal displacements. The last part focuses on the permafrost change and methane flux estimation. Based on the estimated surface deformation from InSAR and methane flux in-situ measurements, we first estimate the degree of changes in permafrost. Then the magnitude of methane flux released from the corresponding area are calculated & estimated on the presumption that there is a linear relationship between the aforementioned two observations. The work-flow of this study is shown in Figure \ref{fig:overview}.
1.4 Thesis layout

- **Chapter 1: Introduction** This chapter gives general background information about the thesis. It describes the motivation, research questions and research methodology of this study.

- **Chapter 2: Permafrost** In this chapter, the focus is on the permafrost nature. The definition and distribution of permafrost are given, followed by its current changes and relative consequences. Then, we summarize the previous permafrost monitoring methods and the current on-going programmes. Specific information about the study area is given afterwards.

- **Chapter 3: Radar Interferometry** This chapter initiates with the principle of Radar Interferometry. The advantages and limitations of this technique are stressed alongside. The principle of the small baseline approach is explained in the last section.

- **Chapter 4: InSAR measurements** This chapter is dedicated to surface deformation mapping. We apply the small baseline approach over the study area to obtain the time-series surface deformation. The processing chain and its corresponding products are explained step-wise.

- **Chapter 5: Deformation modeling** This chapter first explains a deformation model. Then we model the observed InSAR measurements in a Least-square fashion. The quality of modelling is discussed and the modelled results are analyzed afterwards.

- **Chapter 6: Permafrost changes and methane flux** In this chapter, we first estimate the general degree of permafrost changes. Then we describe the assumptions made for the methane flux estimation. The corresponding results are given, including the magnitude of permafrost volume loss and the magnitude of released methane flux.

- **Chapter 7: Conclusion and future work** In this chapter, the thesis is closed with a final conclusion answering the research question. Recommendations about the future works are also included.
Chapter 2

PERMAFROST

In this chapter, we initiate with the background information about permafrost environment, including the definition of different entities and the permafrost distributions. Then, we review the current studies on permafrost changes and its relative consequences. Next, we summarize the previous permafrost monitoring methods and the permafrost on-going programmes. Lastly, we give a general descriptions about the environment in the study area.

2.1 Permafrost environment

2.1.1 Definition

The permafrost layer is defined as “a thermal condition of the ground having a temperature below 0°C for at least two consecutive years” [13]. The permafrost is generally envisaged as frozen sediments and ground ice but it may also consist of the bedrock and organic materials [17]. The formation of permafrost probably dates back to the Pleistocene epoch or even earlier. Permafrost is divided into several zones on the basis of geographic continuity in the landscape. A typical classification recognizes continuous permafrost, underlying 90 – 100% of the landscape; discontinuous permafrost, 50 – 90%; and sporadic permafrost, 0 – 50% [11].

Typically, there are other two distinct layers besides the permafrost layer. The active layer and the unconsolidated sediments layer. The active layer is defined as “a layer of soil or other earth materials lying between the atmosphere and permafrost, which subjects to seasonal freezing and thawing cycles on an annual basis” [13]. It extends from the ground surface to the upper-most of the permafrost layer, called the permafrost table. The active layer thickness (ALT) is a commonly used important factor to infer environmental changes, which usually has a high heterogeneity over space and time. Basically, the ALT depends on the local environment conditions such
as surface temperature, snow cover, vegetation type, terrain slopes, moisture content, thermal conductivity and soil properties [1, 4, 30]. In some places, the unconsolidated sediments layer, also referred to as Talik, is located on top or cutting through masses of permafrost. They are normally wet zones, often with flowing water. See figure 2.1 as an illustration of a typical permafrost environment.

### 2.1.2 Distribution

Typically, permafrost can occur in the areas where mean annual air temperature stands below $-1^\circ$C. In northern hemisphere, regions in which permafrost present occupy approximately 25% of the land area, the estimated permafrost area is about 23 million km$^2$ [21]. Figure 2.2 indicates a global distribution of permafrost in the northern hemisphere.

Most continuous permafrost in the northern hemisphere spread over large parts of Russia, Siberia, north Canada, Alaska and the outer ring of Greenland. In the discontinuous and sporadic zones, permafrost distribution is usually complex and patchy. Permafrost also exists in high mountains, the outstanding examples are present in the rocky Mountain Cordillera as well as Tibetan Plateau. Permafrost extends further south in the eastern hemisphere than in western. The reason might come from the higher elevations southward in eastern hemisphere. Permafrost in middle and low latitude mountains is warm and its distribution is closely related to characteristics of the land surface, such as terrain slope, orientation, vegetation patterns and snow cover. The active layer thickness also increases toward south as the temperature goes warmer.
2.1.3 Current changes

Permafrost under undisturbed terrain most commonly exists in a reasonable state of thermal equilibrium with the climatic environment. However, due to the global warming, the resulting change of the thermal regime beneath the surface could break this equilibrium and cause the permafrost unstable. Recent researches and studies over permafrost area in north Canada, central Alaska, northeast Greenland and north Russia report a common trend in permafrost change. They have found the active layer is thickening in depth and that the permafrost layer is subjected to the thawing process [3, 4, 16, 18, 41, 56, 58, 78]. These observed changes in permafrost environment have considerable impacts on the regional ecosystem as well as the climate system [22, 49, 55].

Since most of the thermal energy exchange is occurring in the active layer, a thickening of the active layer will have profound effects on geomorphic, hydrologic and biological processes [73]. Evidences are usually indicated by the changes of vegetation type and losses of lakes over permafrost environment [12, 15, 67]. The thickening of the active layer and the thinning of the permafrost layer may also mobilize carbon currently trapped in the upper-

Figure 2.2: Northern hemisphere permafrost coverage in purple gradient, classified into four different types of permafrost: Continuous, Discontinuous, Sporadic and Isolated [71].
Table 2.1: Commonly used site-monitoring methods.

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<td>Thickness</td>
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<td>Ground penetrating radar</td>
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most permafrost layer. Thus these changes could trigger a higher emission of greenhouse gases into the atmosphere and pose a positive feedback on the climate change. There are several evidences from different in-situ gas flux measurements over permafrost regions [17, 53, 74]. Particularly, in ice-rich permafrost area, variations in ALT and permafrost thawing both will cause some degree of surface deformation. This deformation over permafrost area has been observed at different sites during the past decades [48, 76].

2.2 Previous permafrost monitoring

2.2.1 Monitoring methods

Until now, in-situ measurement is the prominent way to carry out the permafrost monitoring task. The commonly used methods in site measurements are summarized in Table 2.1. Probing involves pushing a probe rod into the active layer until the solid ground is reached. The depth of the active layer is read from the probe rod at the soil surface. These depth measurements are strictly made relative to the surface [10]. Using Thaw tube, a hole need to be drilled perpendicularly into the frozen ground. Then a PVC outer tube with a clear marked tube inside is inserted. The clear tube is filled with a liquid that changes color when the soil temperature reaches the freezing point and is then sealed at both ends. Then the inner tube is pulled out to estimate the depth of the freezing point. Soil temperature profiles at certain depth can be obtained by using soil probes. Ground Penetrating Radar (GPR) can be used to determine the ice’s thickness. Ice is transparent to microwave radio signals, while ice-sediment and ice-water interfaces are reflective. Therefore, the GPR is effective in the wetland area.

During the past tens of years, site measurements have continuously provided high-quality observations. These measurements have provided valuable information for scientific researches. However, traditional site measurements
are subjected to the poor spatial coverage drawbacks. To better understand the overall permafrost environment, observations covering vast area are highly required.

2.2.2 On-going programmes

**CALM**  One prestigious programme initialized by IPA is the Circumpolar Active Layer Monitoring (CALM) programme. The primary goal is to observe the response of the active layer and near surface permafrost to climate change over long time scales [10]. The CALM monitoring network established in 1990s. CALM currently has participants from 15 countries. Approximately 60 sites measure active layer thickness on grids ranging from 1 ha to 1 km$^2$, and 100 sites observe soil temperatures, including permafrost temperature from boreholes. Figure 2.3 indicates the locations of monitoring sites through the CALM programme.

**ZERO**  Zackenberg Ecological Research Operations (ZERO) is the operating programme in northeast Greenland. The intensive monitoring area is situated at the Zackenberg valley (74°30N, 21°00W) in the national park of Greenland. This monitoring programme was initiated in 1995 and fully implemented for the terrestrial part of the ecosystem in 1996. The objective of this programme is to provide long time series of data on the natural innate oscillations and plasticity of a high arctic ecosystem. The programme was supplemented with five sub-programmes: *ClimateBasis*, *GlacioBasis*, *GeoBasis*, *BioBasis* and *MarineBasis* [44]. The *GeoBasis* sub-programme aims to collect data of hydrological and terrestrial variables including: (1) snow, ice and permafrost; (2) river water discharge and chemistry; (3) precipitation and soil water chemistry; (4) gas fluxes of carbon dioxide and methane and (5) geomorphology [32]. The *GeoBasis* sub-programme has two ZERO-CALM grids (marked by the green box in figure 2.3), the size is about 100x100 m$^2$ of each. Both grids measure the ALT during thawing season of each year. The observed data shows an increasing trend in ALT during the past two decades (refers to APPENDIX A). A degree of carbon dioxide and methane flux has been observed in the same area as well [17]. The changes of the factor, such as increasing in ALT, are important indicators of the local environment situation.

2.3 Area of interest

In Greenland, the ice sheet (1755637 km$^2$) covers 81% of the total area (2166086 km$^2$) [50]. The permafrost occupancy exists around the out ring of Greenland, near the coastline. It’s presence seems quite marginal. Therefore, the Greenland permafrost draws little attention compared to elsewhere.
such as Alaska and north Canada. As far as we know, there is no permanent GPS station near by the location of on-going ZERO programme, this programme does not include a long-term monitoring on the real surface deformation. Therefore, there is an added-value in measuring the surface deformation alongside the ZERO programme. From one side, the long-term monitoring on surface deformation can provide an additional option to study the dynamics inside the permafrost and the active layer. It is complementary to the ZERO programme. On the other side, the ZERO measurements provide useful data which can be used to constrain and validate our remote sensing measurements.

Based on the above added-value, we decide to take the Zackenberg valley inside our study area, as the ZERO programme is running there. Figure 2.4 indicates the location of the study area.

The study area is typical high-arctic tundra zone. The main vegetation types are grassland and fen [17]. The maximum elevation goes up to 1800 meters in this study area. Figure 2.5 shows the landscape near Zackenberg valley. The snow covered mountains, vast tundra area, grassland, rivers, small lakes and spread bared rocks form the typical landscape over there.

Observations from the ZERO programme have shown that the seasonal thawing process usually begins in the middle of May or early June on exposed sites. By the late August or early September, the entire soil starts to refreeze again. See APPENDIX B for the ZERO monitoring data of recorded soil temperature from the year 1996 till 2009.
The seasonal variations in temperature also agree with the surface snow cover periods. From May, when the ground temperature rises above zero, the snow starts to melt. The ground surface stays snow-free through June, July and August. Then the snow comes back in September when the ground temperature drops below zero again. Figure 2.6 shows a camera view of surface-cover variation within one year over Zackenberg valley.

On the basis of observed annual ground temperature variation and the surface snow-cover period, it is safe to say that the thawing season usually starts in May and ends in September of each year over the study area, northeast Greenland.
Figure 2.5: The landscape over the study area, near Zackenberg valley north-east Greenland (Photos by Lars H. Hansen, 2008).

Figure 2.6: A camera view of snow cover variations in the year 2008. The order from the top-left to lower-right corresponds to the month from January till December. Pictures were taken by a digital camera mounted on a prominent rock on the eastern slope of Zackenberg (Provided by ZERO, 2008).
Chapter 3

RADAR INTERFEROMETRY

This chapter aims to review the necessary background information about the Synthetic Aperture Radar Interferometry (InSAR), which is also known as Radar Interferometry. In terms of surface deformation mapping, InSAR has its notable advantages compared to the traditional methods. Here we explain the principle and stress the advantages & limitations of this technique. Relative background information which is important to understand the following chapters is included.

3.1 Conventional InSAR

3.1.1 Synthetic Aperture Radar

Radar is an acronym for Radio Detection and Ranging. The radar antenna emits electromagnetic pulses at a particular microwave wavelength. Typical space-borne radar use wavelengths in the X-band (≈ 3.1 cm), C-band (≈ 5.6 cm), S-band (≈ 10.5 cm), and L-band (≈ 23.6 cm). Radar detects the reflections of these pulses from target. It uses the round-trip time of the pulse to determine the range from target as well as its backscatter intensity to infer target physical properties, such as size and shape. One particular class among the radar systems is the imaging radars, for instance, Synthetic Aperture Radar (SAR). In contrast to traditional radar, which sends directional pulses and detects the presence, position and motion of an object by analyzing the portion of energy reflected from the object, imaging radar attempts to form a picture of the object by mapping the electromagnetic backscattering coefficient onto a two-dimensional plane.

Pixel resolution in the flight direction depends on physical length of antenna. Due to the practical restrictions on the antenna length, the resulting resolution is poor and it is usually in the order of kilometers. In order to im-
prove this resolution, SAR utilizes the movement of the satellite platform to create a 'synthetic' large aperture [33]. The geometry of a SAR acquisition is shown in the figure 3.1.

In sequence, a series of reflected phase signals are received within that SAR image. SAR uses the redundancy in the return signals to reconstruct the scatterers. The strategy used to reconstruct the single scatterer from the series of return signals is referred to as SAR focusing [7, 24, 59, 60, 75].

### 3.1.2 SLC image

After SAR focusing, the product is usually referred to as a Single Look Complex (SLC) image. Satellite SAR missions ERS-1, ERS-2 and ENVISAT are used in this study. They are preprocessed to SLC by one of the European Space Agency Processing and Arching Centers (ESA PACs). The spatial resolution in a SLC image of ESA satellite is about 4 m in azimuth direction.
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<th>Launch Year</th>
<th>ΔT (Day)</th>
<th>$H_{sat}$ (km)</th>
<th>$f_0$ (GHz)</th>
<th>$B_R$ (MHz)</th>
<th>$\theta_{inc}$ (deg)</th>
<th>Swath (km)</th>
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<td>5.3</td>
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<td>20-50</td>
<td>100-500</td>
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</tbody>
</table>

Table 3.1: Important design parameters. $\Delta T$: repeat period; $H_{sat}$: satellite altitude; $f_0$: carrier frequency; $B_R$: range bandwidth; $\theta_{inc}$: incidence angle (Adapted from [33]).

![An example of amplitude (left) and phase (right) retrieved from an ERS SLC image (acquired in 19950510).](image1)

and 20 m in range direction, its operating wavelength is about 5.6 cm (C band), some other important design parameters are listed in Table 3.1.

Each pixel in SLC image is recorded as a complex number. Both amplitude and phase information can be derived from each pixel using equations 3.1 and 3.2.

$$|y| = \sqrt{\{Re\} + \{Im\}}$$

$$\phi = \arctan \frac{\{Im\}}{\{Re\}}$$

Where $|y|$ represents the amplitude of transmitted pulse, $\phi$ represents the phase of signal, $\{Re\}$ indicates the real component of the complex number, $\{Im\}$ is the imaginary component. Figure 3.2 shows an example of amplitude and phase extracted from one of the used SLC images.

The amplitude contains information about the reflective properties of the target, such as the surface roughness, the shape, the orientation. The phase
of signal is a summation of each individual scatterer’s phase responses within that pixel rather than any of the single scatterer.

3.1.3 Interferometry

Interferometry is the technique to infer range difference in phase between two different acquisitions. For convenience, the recorded value for each SLC pixel is written as in the equation (3.3).

\[ y = |y| \cdot e^{(j\phi)} \]  

(3.3)

Where \( y \) is the value of an arbitrary pixel in a SAR image, \( |y| \) represents the amplitude, \( \varphi \) is the phase of signal, and \( j \) is the imaginary operator. Let us consider one pixel in SLC image, which moves away from satellite looking direction. Taken the phase difference indicated in figure 3.3, we can estimate the displacement between the two acquisitions for that pixel.

Taken two satellite acquisitions, usually called Master and Slave, the relative phase difference can be retrieved by a point-wise complex multiplication of one image with the conjugate of another. The product is called an interferogram, which indicates the relative phase difference between corresponding pixels in both images at their acquisition dates. The obtained phase difference contains valuable information such as surface deformation. Before further processing, however, it is only known \( \text{modulo} - 2\pi \), which is referred to as wrapped phase. Equations 3.4, 3.5, and 3.6 illustrate the principle of interferometry.
Where $y_{i,m}$ is the $i^{th}$ pixel in the Master acquisition. $y_{i,s}$ is the corresponding pixel in the Slave acquisition. $^*$ is the conjugate complex operator. The product $\text{If}_g$ is the resulting interferometric phase for the $i^{th}$ pixel. By assigning a color scale to the obtained interferometric phase, the interferograms can be visualized as fringes. Depends on the radar wavelength, for instance, the ERS mission, one full color-cycle change represents 2.8 cm in range difference. Figure [3.4] shows an interferometric product of the study area.

As in the above figure, the fringe contains valuable information. Ideally, considering the same condition during both acquisitions, the only difference is the surface moved in between the two acquisitions, thus the observed phase difference represents the relative surface movement. In reality, however, conditions such as atmosphere and acquisition geometry are changing from time to time. Any of these variations can contribute to the interferometric phase difference in the interferogram. Thus these contributions overprint on the surface deformation signal. These ‘nuisance’ terms resulting from above changes can be removed or mitigated during MT-InSAR processing. The whole processing chain in terms of measuring deformation with InSAR and its relative product will be explained in detail in the chapter [4].

### 3.1.4 Advantages and limitations

Comparing to traditional measurements, such as GPS and leveling, InSAR has the following advantages. For instance, no need for field work, self-illuminating system (working through day and night), weather-independent (still applicable if clouds cover or rain), high spatial coverage (swath wide 100-500 km), and high spatial resolution (tens of meters). However, there are also limitations which might restrict the InSAR application field.

The critical limitation of applying InSAR is decorrelation. Loss of coherence (decorrelation) introduces noise signal in the interferometric phase, thus it makes the interferometric phase less useful. Several driving mechanisms could cause decorrelation. Although the most relevant ones are temporal and geometric decorrelations [77]. The temporal decorrelation is driven by the changing characteristics of scatterers over time, such as the changes in vegetation type, the presence and absence of surface snow cover. The geometric decorrelation is driven by the changing of viewing geometry of...
Figure 3.4: An interferometric (wrapped phase) product in the study area: Zackenberg, the NE Greenland. The fringes represents the difference in phase. This interferogram is generated using two ENVISAT SAR images (Master: 20070626 and Slave: 20070731).

satellite. It results in different local incidence angles. This difference causes same wave-numbers shifted from mapping of the object spectrum onto the data spectra of the two acquisitions [33]. Figure 3.5 illustrates the two kinds of decorrelation problems.

On the upper-left corner of figure 3.5, $B$ represents the real distance between Master and Slave acquisitions. $B_{\text{perp}}$ is the projected distance between Master and Slave, which is perpendicular to the satellite Line-of-Sight (LOS) direction. $B_{\text{perp}}$ is the commonly used term to measure the difference in geometry. This difference in geometry results in two different local incidence angles $\theta_1$ and $\theta_2$. The resulting consequence is shown in the upper-right corner. The data spectrum of two SAR acquisitions shift the same wave numbers in range direction, $N_1$ and $N_2$ are the non-overlapping components in object spectrum between Master and Slave acquisitions. On the lower-left corner, the cartoon illustrates a resolution cell containing several equivalent scatterer elements. Since the return backscatter phase is
Figure 3.5: Two types of decorrelation driving mechanism. The top one represents the geometry decorrelation due to the different viewing geometry between two acquisitions. The lower one represents the temporal decorrelation due to the variations of individual scatterer element within one particular resolution cell over time (Adapted from [33,40]).

A coherent summation of all scatterer elements’ contributions within that resolution cell, any change of each element’s contribution varies the return signal. The lower-right corner shows a simulation of temporal-decorrelation. By varying the phase response from each single scatter elements for 50 times, the received phase responses within the resolution cell is obtained. The phase variation between different acquisitions is quite large, therefore the phase coherence between two acquisitions is quite low. Figure 3.6 shows an example of both decorrelation problems in the study area, the interferometric phase superimposes on the amplitude image.

From figure 3.6, the coherence is low in snow-covered mountain regions, thus the color looks quite noisy and random. While in other snow-free regions, the color looks smooth and spatially correlated. The noise increases as the both baselines goes up, therefore it indicates the decorrelation increased. For ERS and ENVISAT mission, a length of 1100 m perpendicular baseline will cause a total non-overlapping data spectrum in range direction, thus a complete loss of coherence. This length is called the critical perpendicular baseline. Similar to the critical perpendicular baseline, a time interval of about 5 years usually reaches the critical temporal baseline. However, this depends on the expected deformation rate in a particular case.

Besides the decorrelation, there are other limitations of the InSAR technique that restrict its capabilities as well. First, InSAR is a double-difference measurement. It measures the difference between two places at two different acquisition dates. Hence, it cannot measure the absolute surface de-
Figure 3.6: A set of time-series interferograms indicate the decorrelation problem over the study area, Zackenberg NE Greenland. The reasons for these problems are the variations in surface cover types (mainly caused by the snow’s presence and absence) and the increasing view geometry differences. The corresponding temporal and perpendicular baseline for each interferograms are shown. From the left to right, both baselines are increasing while the correlation is decreasing.

formation, but can only determine the relative ground displacements [28]. For instance, if two objects have the same deformation rate towards the same direction, InSAR measurement can only tell there is no difference in movement between these two objects, rather than determining the absolute deformation of each. Second, SAR is a side looking radar. It only measures the one-dimension surface deformation in the radar LOS direction (the range direction). Unfortunately, it cannot solve any relative movement in the satellite flying direction (the azimuth direction).

3.2 Time-series analysis methods

Due to the drawbacks of in-situ measurements, different time-series analysis methods have been developed. The MT-InSAR method involves the simultaneous processing of multiple SAR acquisitions in time. Comparing to the conventional InSAR, this extension provides a time-series analysis on the deformation, thus the evolvement of deformation can be studied. Currently there are two categories of algorithms for processing multiple acquisitions, the persistent scatterer approach [23, 40, 42, 70] and the small baseline approach [8, 40, 61]. Both methods allow selecting the qualified pixels through all interferograms, of which the deformation signal can be extracted. However, they use different strategies to select their qualified pixels.

In general, the persistent scatterer (PS) approach selects pixels which contain a dominant scatterer element, called the PS pixels. These pixels remain quite stable therefore there is less decorrelation over time. This approach has been very successful in the urban area [39]. In contrast to PS
pixel, a pixel contains no single scatterer that dominates others, referred to as the distributed pixel, its decorrelation problem due to phase variation is often large. However, by forming interferograms only with images separated by a short time-interval and a small difference in acquisition geometry, the decorrelation problem is minimized. Decorrelation is further reduced by spectral filtering in range and discarding the non-overlapping doppler frequencies in azimuth.

Hooper defines “pixels whose filtered phase remain stable or decorrelate slowly over short time interval as the target pixel of small baseline approach, called slowly de-correlated filtered phase pixel (SDFP)”. Compared to persistent scatterer approach, small baseline approach takes care of pixels containing no dominant scatterer elements. On the other hand, for PS pixels, filtering in range and azimuth at the cost of resolution may increase the decorrelation problem. Because coarsening the resolution for PS pixel means there is higher chance to include more scatterer elements within that pixel, making the single stable scatterer less pure than before. However, even after filtering, most PS pixels are still qualified as SDFP pixels.

Since the study area is a non-urban area, few anthropological features present there. The distributed pixels are expected to be more than the PS pixels. In order to find out the locations with useable signal, the small baseline approach described by Hooper is applied in this study area. The processing chain of this approach is explained stepwise in the chapter.
Chapter 4

MT-InSAR PROCESSING

For the purpose of time-series surface deformation mapping, we apply the small baseline approach (developed by Hooper [37]) over the study area. The processing chain of this approach is explained stepwise with the corresponding results shown alongside. This chapter is organized in the same order as the processing goes.

4.1 Small baseline differential interferograms

The small baseline approach has been implemented in the extension of StaMPS software [39]. StaMPS stands for Stanford Method for Persistent Scatterer. StaMPS also calls the Delft-oriented Radar Interferometric Software (DORIS) [42] to generate differential interferograms when it needed. Figure 4.1 shows an overview map of small baseline approach workflow.

![Diagram of small baseline approach workflow](image)

Figure 4.1: An overview workflow diagram of small baseline approach. Details in each stage will be unfolded in the relative sections.

Generally, the whole processing chain can be divided into three distinct stages. The section 4.1 explains the process about formation of small baseline
differential interferograms. The section 4.2 describes the coherent pixels for small baseline approach and the selecting process. The last section, the section 4.3 describes the strategies applied on the selected pixels to separate surface deformation signal among other nuisance terms.

4.1.1 Datasets and master selection

In this study, the processing starts with a dataset of focused SAR images, otherwise called the SLC products. We use both ERS SAR and ENVISAT ASAR collections to cover the period over last two decades, which span from 1992 till 2009. Table 4.1 lists the used images in this study.

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Table 4.1: A list of used acquisitions used in this study. The types of sensor and the orbit number are shown. The frame number is 2087. The track number is 439 for ERS and 210 for ENVISAT. Dates of acquisitions are in format yyyymmdd.

In total, 23 images are used. However, we treat them as two subsets based on their types of mission: the *ERS* subset and the *ENVISAT* subset. It is important at this point to address that all the images used here are
acquired within the period from May to October. This covers the period of thawing season which we have already mentioned in chapter 2.3. Since outside the thawing season, the vast area is covered by snow. InSAR is not applicable in snow-covered area due to the decorrelation problems as indicated in figure 3.3.

As the small baseline approach is implemented in StaMPS, initially, a master image needs to be chosen. Then all other slave images are co-registered and re-sampled with respective to the single master. The small baseline interferograms are formed afterwards by recombinating the re-sampled SLC images. Further, the differential interferograms can be obtained by subtracting the geometric and topographical phase contributions. Figure 4.2 shows the processing chain to generate the small baseline differential interferograms in StaMPS.

![Flow diagram of small baseline interferometric process in StaMPS (Adapted from [42]).](image)

The goal of master selection is to minimize the total decorrelation of all the interferograms. A model includes a product of four terms is given
by Hooper [39] in equations 4.1, 4.2 including the time-interval \((T)\), the perpendicular baseline \((B \perp)\), the difference in doppler centroid \((F_{DC})\) and the thermal noise [77].

\[
\rho_{total} = \rho_{temporal} \cdot \rho_{spatial} \cdot \rho_{doppler} \cdot \rho_{thermal} \quad (4.1)
\]

\[
\approx \frac{1}{N} \sum [1 - f(\frac{T}{T_c})][1 - f(\frac{B \perp}{B_{c \perp}})][1 - f(\frac{F_{DC}}{F_{DC}^c})] \rho_{thermal} \quad (4.2)
\]

Where \(N\) is the number of image-pairs, \(\rho\) denotes the correlation, and the superscript \(^c\) means the critical value of factor. The function \(f(x)\) is given in equation 4.3.

\[
f(x) = \begin{cases} 
  x, x \leq 1 \\
  1, x > 1 
\end{cases} \quad (4.3)
\]

For \textit{ERS} and \textit{ENVISAT} mission. The typical critical values are \(T_c = 5\) yr, \(B_{c \perp} = 1100\) m and \(F_{DC}^c = 1380\) Hz. Based on this model, the selected master and estimated average correlation for both subsets are shown in the table 4.2.

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Table 4.2: A table of master selection results. The images in Bold are the selected masters (with maximum \(\rho\) among others) for both subsets respectively.

StaMPS has the capability to process multiple image-pairs (stack) simultaneously. In the single pair process, coregistration between images with large baselines may lead to failure. The multiple image-pairs process provides opportunities to avoid co-registering the slaves with large perpendicular baselines directly to the master image. Instead, the slave is co-registered to another ‘assisting’ slave, which has smaller perpendicular baselines both
to the previous slave and to the master. Thus it assists the co-registration procedure, made the co-registration more reliable. Once the coregistration has been done, all slave images have directly or in-directly aligned themselves to the master image. Then, all slaves are resampled onto the same sampling girds of the master image. After all images have been aligned and resampled to the same grids, any arbitrary image recombinations from all these images are ready for the point-wise multiplications.

4.1.2 Small baseline network construction

In order to maximize the correlation, small baseline approach aims to minimize the temporal, the perpendicular and the doppler baselines when interferograms are formed. However, in this particular case, the expected deformation is long-term based and fairly slow. Therefore some long temporal baseline is needed. For most cases, the Doppler separation is well below the critical value. Therefore priority is given to the length of the perpendicular baseline to minimize the geometric decorrelation. Table 4.3 lists the small baseline image re-combinations founded in this study.

There are 57 image combinations founded for small baseline approach. Most of which have perpendicular baseline smaller than 200 m, with a maximum value of 408 m. The temporal baselines vary from a minimum 1 month to 4 years at most.

4.1.3 Differential interferograms formation

Before interferometry processing, noise can be reduced by first applying the spectral filters in azimuth direction and then in range direction. Filtering in azimuth is to exclude non-overlapping doppler spectrum [37]. Filtering in range is to cut off the shift component mapping from object spectrum to data spectrum [33]. However, both filters are slightly different in strategies [33, 42].

After filtering in both directions, small baseline interferograms are formed for all the founded image-combinations. The products are wrapped interferograms. Due to different acquisition geometry, each interferogram contains phase contribution from the reference surface (WGS-84 for now) and the surface topography. These two terms can be computed and subtracted from the wrapped interferometric phase respectively. Once the precise orbit information is available, the reference phase is simulated from the orbit ephemeris. Simulation of the topographical phase needs an external digital elevation model.

In this case, the Delft orbit products are applied to compute the reference phase. These precise orbit files provide the satellite ephemeris information (latitude, longitude and height) every 60 seconds. The files were generated by the Delft Institute for Earth-Oriented Space Research (DEOS). The
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Table 4.3: A list of used small baseline image re-combinations for InSAR processing.
ASTER GDEM is used to simulate the topographical phase contribution. Figure 4.3 shows the picture of ASTER GDEM covered the study area. The ASTER GDEM is in WGS-84 system and it is in an equi-angular grid with 1 arc-second posting interval (about 30 meters). The format of this DEM is short integer.

However, in some places, ASTER GDEM does contain residual anomalies and artifacts that degrade its overall accuracy. The accuracy standard deviation is declaimed as 7-14 m \(^5\), while in mountain area this estimation might be even worse. By generating profiles in the east-west direction, the anomalies are identified. Figure 4.4 enumerates several profiles where the anomalies value are found in the study area.

A 41x41 window size median filter has been applied to the specific anomalies region in the original ASTER GDEM source. The filtered results are as expected: the anomalies are filtered out while the good samplings are preserved. Figure 4.5 compares the prior profiles with the filtered ones.

This filtered DEM is used in simulation of topographical phase contribution. First, the DEM is coded to the radar coordinate systems of the master image. Then the phase is interpolated to integer grids of master coordinates. A linear interpolation based on a Delaunay triangulation is used \(^55\).\(^66\).

Then, we subtract both estimated phase contributions from interferograms by complex multiplications with their conjugate patterns. Equation 4.4 is a simple model subtracting the corresponding terms and obtaining the differential interferograms.

---

Figure 4.3: The external DEM source: ASTER GDEM (study area indicated by red square box).
Figure 4.4: ASTER GDEM profiles. The left side red line indicates the location of profiles being generated. The right side shows the corresponding profiles. In each of above listed profiles, punch-like signals are not realistic at all, they are treated as anomalies in the source.

Figure 4.5: Median filtering. The original profiles are on the left side. The filtered ones are on the right side. Punch-like signals are filtered out.
\[ I = M \cdot S^* \cdot R^* \cdot T^* \]  \hspace{1cm} (4.4)

Where * denotes the complex conjugate operator. \( I \) is the complex differential interferogram. \( M \) is the complex master image. \( S \) is the complex resampled slave image. \( R \) is the complex reference phase. \( T \) is the complex topographical phase. Figure 4.6(a) and 4.6(b) shows all the obtained small baseline differential interferograms.

The left phase in each interferograms is still being wrapped. Therefore it is still difficult to draw any conclusion from the wrapped products. At this point, however, it is more important that all interferograms should remain some degree of coherence, because it will influence the following process in pixel selection.

### 4.2 Coherent pixel selection

The definition of coherent pixel for small baseline approach has been given by Hooper [37] (refers to chapter 3.2). The identification of these coherent pixels uses the same algorithm as described by Hooper [39]. Primarily, this algorithm uses the correlation of phase in space to identify the coherent pixels rather than the variation of phase over time. For the reason of computational load reduction, the amplitude characteristic is analyzed at first. A measure \( D_{\Delta A} \) is defined and it acts as an indicator of potential coherent pixel phase stability [37]. The definition of \( D_{\Delta A} \) is given in equation 4.5.

\[ D_{\Delta A} = \frac{\sigma_{\Delta A}}{\mu_A} \]  \hspace{1cm} (4.5)

Where \( D_{\Delta A} \) is the amplitude difference dispersion. \( \sigma_{\Delta A} \) is the standard deviation of the amplitude difference between each pair. \( \mu_A \) is the mean amplitude. A reasonable value \( D_{\Delta A} = 0.6 \) is used for the computational load reduction [37]. The corresponding subset which referred to as coherent pixel candidates. It contains most of potential coherent pixels, but also including pixels that are false-positive. Further, the phase characteristic is used to find out real coherent pixels among all candidate pixels.

At this moment, the phase contributions in the interferograms includes the following terms as indicated in equation 4.6 [39].

\[ \varphi_{\text{int}} = W\{\phi_{\text{def}} + \phi_{\text{atm}} + \Delta\phi_{\text{orb}} + \Delta\phi_{\theta} + \phi_n\} \]  \hspace{1cm} (4.6)

Where \( \varphi_{\text{int}} \) is the wrapped interferometric phase. \( \phi_{\text{def}} \) is the phase change due to movement of pixel in the satellite LOS direction. \( \phi_{\text{atm}} \) is the phase change due to atmospheric delay. \( \Delta\phi_{\text{orb}} \) is the residual phase due to orbit
Figure 4.6: The obtained differential interferograms for both subsets. The reference and topographical phase contributions have been subtracted from the interferometric phase. The upper ones are obtained from ENVISAT subset; the lower ones are from ERS subset.
inaccuracy. $\Delta \phi_{\theta}$ is the residual phase due to look angle error. $\phi_n$ is the phase due to all other noise terms and $W\{\}$ is the wrap operator.

$\phi_n$ is the term we want to have an access. Pixel having smaller $\phi_n$ indicates less decorrelation thus higher coherence in interferograms. Therefore the first four terms in equation 4.6 need to be estimated and temporally subtracted from the wrapped phase. Hooper [39] treats the first four terms as two groups: the spatial-correlated terms and the spatial-uncorrelated terms.

The spatial-correlated contributions of one single pixel to the interferometric phase is estimated by band-pass filtering of mean neighboring pixels. This is assumed to include the phase change due to surface displacement, atmospheric delay, orbit inaccuracy and spatial-correlated part of look angle error. See equation 4.7.

$$\tilde{\varphi}_{\text{int}} = W\{\hat{\phi}_{\text{def}} + \hat{\phi}_{\text{atm}} + \Delta \hat{\phi}_{\text{orb}} + \Delta \hat{\phi}_{\theta}^{sc}\}$$ (4.7)

Where $\tilde{\varphi}_{\text{int}}$ is the estimated spatial-correlated part of $\varphi_{\text{int}}$. $\Delta \hat{\phi}_{\theta}^{sc}$ is the estimated spatial-correlated look angle (SCLA) error. $\Delta \hat{\phi}_{\text{orb}}$ denotes the estimated error due to orbit in-accuracy. $\Delta \hat{\phi}_{\text{atm}}$ indicates the estimated atmospheric delay. $\Delta \hat{\phi}_{\text{def}}$ denotes the estimated surface deformation.

The spatial-uncorrelated look angle (SULA) error includes contributions from both spatial-uncorrelated DEM error and deviation of the pixel’s phase center from its physical center. It is estimated though its correlation with perpendicular baseline [39] as in equation 4.8.

$$\Delta \hat{\phi}_{\theta}^{u} = \frac{4\pi}{\lambda} B_{\perp} \Delta \hat{\theta}^{u}$$ (4.8)

Where $\Delta \hat{\phi}_{\theta}^{u}$ is the spatial-uncorrelated look angle error. $\lambda$ is the wavelength. $B_{\perp}$ is the perpendicular component of the baseline. $\Delta \theta^{u}$ is the error in look angle. Subtraction of both estimated terms leaves the de-correlation noise term for that pixel, the phase contribution equation can be rewritten as:

$$\varphi_{\text{int}} - \tilde{\varphi}_{\text{int}} - \Delta \hat{\phi}_{\theta}^{u} = W\{\phi_n + \delta\}$$ (4.9)

Where $\delta$ includes all estimated residues which is expected to be small. Hence the left term $\phi_n$ dominates the residual term. Figure 4.7(a) 4.7(b) 4.7(c) shows one estimated result of above three terms in our case: (a) $\varphi_{\text{int}}$, (b) $\tilde{\varphi}_{\text{int}}$ and (c) $\Delta \hat{\phi}_{\theta}^{u}$ respectively.

Then a measure similar to coherence magnitude is defined by Hooper [39] as in equation 4.10.
(a) Wrapped phase (unit: rad).

(b) Estimated spatial correlated terms (unit: rad).

(c) Estimated spatial un-correlated look angle error (unit: rad).

Figure 4.7: The estimated three terms in equation 4.9.
\[
\gamma_x = \frac{1}{N} \left| \sum_{i=1}^{N} \exp\{j(\varphi_{\text{int}} - \tilde{\varphi}_{\text{int}} - \Delta\hat{\phi}_u^i)\} \right|
\] (4.10)

Where \(N\) is the number of interferograms. Statistical analysis on the distribution of \(\gamma_x\) and \(D_{\Delta A}\) yields a threshold function, which depends on a specified value for the required certainty that the phase of a selected pixel is not random [39]. In this case, there are 110037 coherent pixels found through ERS subset and 139542 coherent pixels found through ENVISAT subset (Pixels have been resampled to a 100x100 m\(^2\) grid for the reason of computational load).

### 4.3 3D phase unwrapping and nuisance terms

Phase unwrapping is the process of recovering unambiguous phase values from phase data that are measured modulo 2\(\pi\) rad. In the small baseline approach, the unwrapping starts after coherent pixels selected. The unwrapping process is based on the Nyquist criteria, which assumes that the phase difference between neighboring sample points in any dimension is generally less than half a phase cycle. In other words, accurate phase unwrapping is only possible when the absolute difference in phase between neighboring points is less than \(\pi\) rad, otherwise a situation called phase discontinuity occurs. It is usually assumed that the spatial sampling rate is high enough over most of the area, even though, this criterion only holds for the spatial-correlated terms but not for the spatial uncorrelated term [36]. Therefore, the spatial uncorrelated term is subtracted from the interferometric phase before phase unwrapping. The left terms are more optimal for the unwrapping process.

Here the 3D phase unwrapping method (developed by Hooper [39]) has been applied in this study. Compared to conventional 2D unwrapping in space domain, 3D unwrapping improves the solution in a similar way to which 2D unwrapping provides an improvement over the 1D method. First, it provides more potential paths bypassing any phase discontinuous regions. Second, the presence of residue provides clues to identify these locations of phase discontinuity regions. A residue is a point around which integrating the phase gradient does not return zero [29]. Figure 4.8 shows the relative 3D unwrapping routine for small baseline approach implemented in StaMPS software.

The sparse neighboring pixels are interpolated in spatial domain using nearest-neighborhood algorithm. The interpolated results are resampled to the regular grids. Phase difference between neighboring grids is calculated for all interferograms. The wrapped phase are first applied a low-pass filter in time using a Gaussian weighted window and then being unwrapped in
temporal domain under the Nyquist assumption. A prior probability density functions (PDFs) is formed for the phase difference between the neighboring grids in each interferogram. Cost functions are derived from the prior PDFs by taking the negative logarithm. Then, the optimization routines of SNAPHUB [38] are used in spatial domain to search for the minimum total cost solution for each interferograms [51].

Till now, the phase contribution in the retrieved single master interferograms can be rewritten as indicated in equation 4.11.

$$U_w\{\varphi_{\text{int}} - \hat{\Delta}^u\} = \phi_{\text{def}} + \phi^M_{\text{atm}} + \phi^S_{\text{atm}} + \Delta \phi_{\text{orb}} + \Delta \hat{\phi}_{sc} + \hat{\phi}_n + \delta \quad (4.11)$$

Where $U_w\{\}$ is the unwrapping operator. $\Delta \hat{\phi}_{sc}^u$ is the estimated spatial-uncorrelated term, which is subtracted from wrapped phases before unwrapping. On the right side of equation, $\phi_{\text{def}}$ is the deformation signal. $\phi^M_{\text{atm}}$ and $\phi^S_{\text{atm}}$ are the Master and Slave atmospheric delay respectively, mainly due to the change of water vapor content in troposphere along paths between acquisitions. The master atmospheric delay and the orbit error $\Delta \phi_{\text{orb}}$, together called Atmosphere & Orbit Error (AOE), present though all interferograms. They are estimated simultaneously. The slave atmosphere delay $\phi^S_{\text{atm}}$ present differently in each interferogram which has been estimated individually. The SCLA error $\Delta \hat{\phi}_{sc}^c$ is almost exclusively due to spatial-correlated DEM error. In StaMPS, the unwrapping process and nuisance term estimations are integrated in an iterative loop. This iteration aims to refine the estimation of SCLA error and AOE, thus it improves the reliability of phase unwrapping as well. Figure 4.9 illustrates the flow diagram of the iterative process.

Among small baseline interferograms, the time intervals are usually arbitrary and overlapped. This is not straightforward to see the evolution of deformation over time. Therefore, an additional step is applied to retrieve the time-series signal with respective to the single master image. This is done by a weighted least-squares inversion, with a full variance-covariance matrix estimated from the spatial coherence [37]. Then, in order to check
Figure 4.9: Flow diagram of iterative unwrapping and nuisance terms estimation. Items in the green parallelogram boxes represent products. The rectangle box represents process and action. The diamond represents the user interaction. The main routine is indicated by a **bold** arrow chain. The loop starts from the Yellow box.

Whether the phase for all small baseline interferograms contributing to each single master interferogram is consistent or not, the residues that between the small baseline interferograms and the predicted model value from retrieved single master interferograms are calculated. Isolated residue up to $2\pi$ is generally accepted due to the local phase-unwrapping error. However, any spatial-correlated residues imply systematic phase-unwrapping errors. Wrongly unwrapped interferograms can be dropped either prior to the nuisance terms estimation or prior to 3D unwrapping process. Usually, by temporally subtracting the estimated SCLA error from the wrapped phase before unwrapping improves the reliability of unwrapping results. An extra processing to calculate the SCLA error, AOE for single master interferograms is carried out alongside. This iterative process stops till all interferograms have been reliably unwrapped, which are indicated by the estimated residue map.

In this case, all interferograms that have been found in table 4.3 are used in this loop. After three times iterations, the reliability of unwrapping has been improved a lot. Figures 4.10(a), 4.11(a) show the residue evolution during the iterations.
The spatial-correlated residues present in the first time iteration. However, by temporally subtracting the estimated SCLA error and SULA error before unwrapping in the next iteration, the reliable of unwrapping improves. In this case, after re-running the process for three times, the spatial correlated residues are gone only left the individual residuals less than $\pi$, which is acceptable due to the local unwrapping error. The estimation of nuisance terms also benefit in this iterative process.

Figure 4.12(a) shows the reliably unwrapped interferograms, the unwrapped phase contains all the terms expressed on the right side in equation 4.11. Figures 4.13(a), 4.14(a), 4.15(a) and 4.15(b) indicate the estimated nuisance terms respectively.
Figure 4.12: Reliably unwrapped interferograms. Phase contains contribution from deformation and all other nuisance terms: SCLA error, AOE, Slaves atmosphere delay. The unit is (rad).

Figure 4.13: The SCLA error includes the DEM height error itself and the inaccurate mapping error to radar system. It is estimated using the relation with perpendicular baselines. The left one is for ERS subset; the right one is for ENVISAT subset. The unit is (rad/m).
Figure 4.14: The estimated master AOE contribution. The left one is for ERS subset; the right one is for ENVISAT subset. The unit is (rad).
Figure 4.15: The estimated slave atmosphere delay for each interferograms. This atmosphere delay (mainly due to water vapor change along the path of radar signal in the troposphere) is estimated using the relation between phase and elevation [23]. The unit is (rad).
We have already estimated all the nuisance terms including the SCLA error, the master AOE, and the slave atmospheric delay. Subtracting all these nuisance terms from the unwrapped interferograms. The phase contribution equation can be rewritten as indicated in equation 4.12.

\[ \phi_{\text{int}} - \Delta \hat{\phi}_u^\theta - \hat{\phi}_{\text{atm}}^M - \hat{\phi}_{\text{atm}}^S - \Delta \hat{\phi}_{\text{orb}} - \Delta \hat{\phi}_{\text{sc}}^\theta = \phi_{\text{def}} + \hat{\phi}_n + \delta \]  (4.12)

Where \( \phi_{\text{int}} \) denotes the unwrapped phase in one interferogram. \( \Delta \hat{\phi}_u^\theta \) is the estimated spatial un-correlated look angle error. \( \hat{\phi}_{\text{atm}}^M \) and \( \hat{\phi}_{\text{atm}}^S \) denote the estimated atmospheric delay. \( \Delta \hat{\phi}_{\text{orb}} \) is the estimated orbit error. \( \Delta \hat{\phi}_{\text{sc}}^\theta \) is the estimated spatial-correlated look angle error. \( \hat{\phi}_{\text{def}} \) is the phase change due to surface displacement in the radar LOS direction. \( \hat{\phi}_n \) is the estimated decorrelation noise term. \( \delta \) includes all residuals in estimations. Since we only focus on the coherent pixels, of which the decorrelation noise is quite small. In other words, on the right side of equation 4.12, the deformation signal dominates the noise term and the residues. Figure 4.16(a) 4.16(b) shows the phase contribution mainly due to deformation.
Figure 4.16: The phase contribution is mainly due to surface deformation. The upper one (a) obtained from ERS subset; the lower one (b) is from ENVISAT subset. The unit is (rad).
The left phase contribution is now mainly caused by deformation. We have converted the left phase (in the unit rad) into deformation (in the unit mm) using the relation given by equation 4.13:

$$defo = -\frac{\lambda}{4\pi}\phi$$  \hspace{1cm} (4.13)

Where $\lambda$ is the radar wavelength (ERS: 56.6 mm, ENVISAT: 56.2 mm). $\phi$ is the phase due to surface deformation. $defo$ is the deformation in millimeter. These obtained time-series deformation will be used in chapter 5 as our MT-InSAR observations.
Chapter 5

DEFORMATION MODELING

In this chapter, the observed deformation from MT-InSAR will be studied. We use a deformation model to investigate the observed surface deformation. The results from this model are interpreted and analyzed afterwards. Conclusions about the surface deformation relative to permafrost change are drawn in this chapter.

5.1 Deformation model

Based on the current observed permafrost changes (refers to chapter 2.1.3), the following model is involved to study the deformation over permafrost area as indicated in equation 5.1. Since we only use SAR images acquired in summer, this model is used to express the deformation within a thawing period during the summer time.

\[ D = R \cdot (t_2 - t_1) + A \cdot (\sqrt{t_2 - t_{1,thaw}} - \sqrt{t_1 - t_{1,thaw}}) \] (5.1)

Where \( D \) is the MT-InSAR observed relative deformation. \( t_1 \) and \( t_2 \) are two SAR acquisition dates respectively. \( t_{1,thaw} \) and \( t_{2,thaw} \) are the onset dates of the thawing season in that year. \( R \) is the long-term deformation rate. \( A \) is the seasonal variation coefficient.

Basically, the first component \([R \cdot (t_2 - t_1)]\) is a linear term which describes the long-term surface deformation caused by thawing in the permafrost layer. The second component \([A \cdot (\sqrt{t_2 - t_{1,thaw}} - \sqrt{t_1 - t_{1,thaw}})]\) simulates the surface deformation due to the snow depth evolution in the active layer during one thawing season. The square-root-of-thawing-days relation is based on the simplified Stefan equation [25]. It says in particular that the expected ice accretion is proportional to the square root of the number of degree
Figure 5.1: Simulation of the surface deformation model, x axis is the time in the unit of days. y axis is the relative deformation. The upper one is the simulation of active layer seasonal variations in snow depth within thawing seasons, each solid curve represents the thawing season within that year and the dash curve indicates the refreezing season in that year. The dash line shown here is just simply connected with the two neighboring thawing seasons, it is not the estimated value from this model. The middle one simulates the long-term deformation due to permafrost thawing underneath. The lower one is the mix of these two signals.

days below freezing and vice versa. The Stefan equation is commonly used in analytic estimates of active layer thawing depth [58]. In this study, the simplified Stefan equation is used to simulate the surface deformation caused by snow depth evolution in its melting process within the thawing season. In other words, this model is not used to estimate the surface deformation during the refreezing phase. Figure 5.1 shows a simulation from this model.

In this study, in order to make the model more flexible, we have added a constant parameter to this model. Thus, there are three unknown parameters in the model: the long-term deformation rate $R$, the amplitude coefficient $A$ and the constant $C$ as in equation 5.2.

$$Defo = R \cdot (t_2 - t_1) + A \cdot \left( \sqrt{t_2 - t_{2,\text{thaw}}} - \sqrt{t_1 - t_{1,\text{thaw}}} \right) + C \quad (5.2)$$

Generally, on the basis of monitored temperature profile (see chapter 2.3), the thawing period is fixed as it is from 15$^{th}$ May till 15$^{th}$ September. Hence the whole thawing season is about 120 days in length within one year.
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Table 5.1: A table of InSAR time-series measurements, the text in bold are the used observations for the deformation modeling.

Then, the unknown parameters in the model is estimated in a least-square fashion. The inputs are the MT-InSAR deformation observations (in millimeter). Table 5.1 lists the used acquisitions for the deformation modeling.

For the unused observations in the deformation modeling, they are either outside the defined thawing season (such as 20041019, 20081028) or they have large temporal gaps in between (such as the gap between 19920914 and 19950614; 20041019 and 20060711). The least-square estimation is constructed using equation 5.3.

\[ Y = MX \quad (5.3) \]

Where \( Y \) is the vector of used InSAR observations, \( M \) is the design matrix and \( X \) contains the unknown parameters. They are built as in the following way:

\[ Y = \begin{bmatrix} y_i \\ \vdots \end{bmatrix} \quad (5.4) \]

\[ M = \begin{bmatrix} t_2^i - t_1^i & \sqrt{t_2^i - t_{st.thaw}} - \sqrt{t_1^i - t_{st.thaw}} \\ \vdots & \vdots \\ \vdots & \vdots \end{bmatrix} \quad (5.5) \]

\[ X = \begin{bmatrix} R \\ A \\ C \end{bmatrix} \quad (5.6) \]
$y_i$ is the $i^{th}$ used observations in the above table (text in bold). The unknown parameters are estimated using equation (5.7). The residues between the observations and estimations are calculated using equations (5.8),(5.9). The estimated $\hat{R}$, $\hat{A}$, $\hat{C}$ and $\hat{res}$ are stored as matrix for following analysis.

\[
\hat{X} = \frac{M^T \cdot Y}{M^T \cdot M} \quad (5.7)
\]

\[
\hat{Y} = M \cdot \hat{X} \quad (5.8)
\]

\[
\hat{res} = Y - \hat{Y} \quad (5.9)
\]

### 5.2 Quality check

We use the root-mean-square (RMS) of residues to indicates the quality of the modelling. In mathematics, the RMS is a statistical measure of the magnitude of a varying quantity. Therefore, we expect that smaller RMS means more stable in modelling compared with a larger RMS value. For each pixel, the estimated residues are a set of values \{\hat{res}_1, \hat{res}_2, \hat{res}_3, ..., \hat{res}_n\}. The RMS is calculated for each pixel as in the equation (5.10).

\[
RM\hat{S}_{res} = \sqrt{\frac{\hat{res}_1^2 + \hat{res}_2^2 + \hat{res}_3^2 + ... + \hat{res}_n^2}{n}} \quad (5.10)
\]

Where $n$ is the number of used observations in this modeling. The estimated $RM\hat{S}$ together with its histograms are visualized in the figure (5.2).
Figure 5.2: The estimated magnitude of RMS for both subsets. According to the histogram, the ERS subset has a mean value of 2.451, the ENVISAT subset has a mean value of 1.588.

In general, most of the estimated RMS are in the magnitude of 0–7. The ENVISAT subset has been better estimated compared with the ERS subset according to the RMS level. However, The RMS maps only give a general feeling about the overview estimated quality. In order to make the quality of estimation more clear, we have checked the point-wise profile for different pixels to show how the different magnitude of RMS indicates the quality of estimation. As expected, there are four types of estimated results. Basically, the good estimated results can be distinguished from the bad ones, which forms two groups of estimations. See figure 5.3.
Figure 5.3: The situation in Group I (long-term subsidence) got well-estimated using this model. However, situations in Group II (others) got poorly-estimated by this model. The $R$ is the long-term surface deformation rate. The $A$ is the seasonal variation coefficient.

In the figures 5.4(a) and 5.5(a), the pixel-wise profiles with different RMS magnitude are shown for well-estimated situations (Situation A in Group I). The red crosses represent the used observations in modelling, while the green ones are the unused observations because they are outside the assumed thawing season.

From the pixel-wise profiles, it is clear to see how the magnitude of RMS indicates the estimated quality. In situation A, even when the RMS goes up to the magnitude of 7, the model can still fit the observed trend in a good manner. The reason for a larger RMS might be the variation of the onset thawing date within the certain year. For instance, If we look the subfigures 5.5(b) and 5.5(c) in the thawing season of the year 2006, the fitted value stands on the right side of the observations, which may indicate that the actual onset of thawing date starts earlier than our assumption. Similarly, in the year of 2008, the actual onset thawing date probably starts later than we assumed. Because we have assumed the fixed starting date for each year, the model will be probably not going through the observations if there is any advance or delay in the onset thawing date of that year. Therefore the RMS goes up. However, this variation does not change the potential trend of the observations. The model still catch the deformation trend and seasonal variation quite well. Hence it doesn’t contaminate the estimations in terms of long-term deformation rate and seasonal variation scale.

For the situations in Group II, the observations are not properly fit by the model by checking the time-series profile. Even with a small magnitude of RMS value, the fitting goes quite poor. Figure 5.6(a) and 5.6(b) and 5.6(c) shows examples for situations in Group II.

For these three situations, the poor-fit might be reasonable. Because the observations themselves do not look like caused by permafrost thawing. For instance, within the thawing season, some of the above observations show up-lift signal, which goes to the opposite direction as we expected. One possible reason for the up-lift signal during thawing season might be due to the post-glacial rebound. Hydrological loading and unloading can also causes uplift signal. Since these observations looks quite scattered, it is quite difficult to conclude these observations with a simple linear trend.
Figure 5.4: Time-series profiles for situation A with different magnitude of the RMS: from 0 to 3.
Figure 5.5: Time-series profiles for situation A with different magnitude of the RMS: from 4 to 7.
Figure 5.6: Time-series profiles for situation B, C and D in Group II.
More observations covering longer time scale is required in order to find out whether a real up-lift trend exist. More investigations need to be done before the particular driving mechanism for the up-lift signal can be answered.

According to our objective in terms of the methane fluxes estimation. In the following study, we only focus on the locations where the model value indicates a long-term subsidence rate (as situation A in Group I). Because in the ‘up-lift’ area, we do not expect there is permafrost under long-term thawing procedure, no methane flux is expected to be released from permafrost in that area either.

5.3 Deformation analysis

For those locations where long-term subsidence was found, the analysis is going on in the following two aspects. One is the long-term subsidence rate (the estimated $\hat{R}$) and another is the seasonal variation coefficient (the estimated $\hat{A}$).

5.3.1 Long-term subsidence rate

The long-term subsidence rate is the parameter $R$ in the deformation model. By visualizing the estimated values of $R$, the subsidence rate map for both periods over the study area are shown in figure 5.7(a), 5.7(b) (the negative sign represents subsidence in all following plots). The reference area is selected at location $[74^\circ 24.6' N, 21^\circ 60.0' W]$. It situates on a top of ice-free mountain area. It is selected because this area is expected to be more stable compared with wetland or tundra area during the two periods. By comparing the two maps, we found that the subsidence rate has been slightly increased. This change can be clearly seen through the histogram statistics for both periods in figure 5.8.

According to the histograms for both periods, the mode has shifted approximately from -0.4 mm/yr to -1.8 mm/yr. During the period 1995-1999, the mean and standard deviation value of thawing rate is $-1.3 \pm 1.0$ mm/yr. When it comes to the period 2006-2009, the value goes up to $-1.8 \pm 0.9$ mm/yr. In most subsidence locations, the magnitude of subsidence rate is within 0-5 mm/yr. The observed long-term surface subsidence indicates the thawing in the permafrost layer, which we will have an estimation in the chapter 6.

Further, an acceleration map is calculated using $a = \frac{\Delta v}{\Delta t}$. This map indicates the changes in subsidence rate during the two periods. The map and histogram are shown in figure 5.9(a). Based on the statistics, during these 10 years interval over vast subsidence area, the change of subsidence rate varies within a magnitude of 0.2 mm/yr$^2$, with a mean and standard deviation about $-0.04 \pm 0.31$ mm/yr$^2$. Therefore, The subsidence rate remains quite stable during the 10 years interval. On the other side, it is proper to
use the simple linear parameter $R$ to model this long-term subsidence rate within each two periods, since the time interval in each periods is shorter, the variation in rate is even smaller.
Figure 5.7: Long-term subsidence rate for both periods. The negative sign represents subsidence. The unit is mm/yr. The color bars are in the same scale. Both plots are referred to the same area, indicated by the white △. (picture rotated 90° clockwise).
Figure 5.8: Statistic histograms of long-term subsidence rate in both periods. The blue one is for the periods 1995-1999. The red one is for the periods 2006-2009. The negative sign represents subsidence. The unit is mm/yr.
Figure 5.9: The acceleration in subsidence rate, from 1995-1999 to 2006-2009. The negative sign represents accelerating in subsidence rate (picture rotated 90° clockwise).
5.3.2 surface seasonal variations

In this study, the thawing season period is fixed which starts from the 15\textsuperscript{th} of May till 15\textsuperscript{th} of September for each year. Therefore the length of one full thawing season is around 120 days. In order to access the magnitude of this full seasonal variation when it reaches the maximum development, the full thawing period 120 days is used to investigate the full seasonal variation scale. The magnitude of seasonal variation is obtained using following equation 5.11. The results are shown in figures 5.10(a), 5.10(b) and 5.11(a).

\[
\hat{S}_v = \hat{A} \cdot \left( \sqrt{t_2 - t_2\text{thaw}} - \sqrt{t_1 - t_1\text{thaw}} \right) = \hat{A} \cdot \sqrt{120} \tag{5.11}
\]

Where \(\hat{S}_v\) is the estimated surface seasonal variation scale, \(\hat{A}\) is the seasonal variation coefficient, which we already have an estimation.

According to the statistics, the mode of histogram has shifted approximately from -1 mm to -5 mm. During the period 1995-1999, the magnitude of variation is about 0–10 mm within one thawing season, the mean and standard deviation value is \(-3.2\pm2.8\) mm. During the period 2006-2009, the magnitude relocates to 0–15 mm. The mean and standard deviation value is \(-5.1\pm2.8\) mm in one full thawing season. A difference map is shown in figure 5.12(a), 5.12(b) to indicates the changes in seasonal variation scale between these two periods.

In terms of changes in seasonal variation scale, the statistics show that most subsidence area is under an increment in surface seasonal variation scale with a mean and standard deviation value of \(-1.9\pm3.7\) mm between the two periods. The observed increment in surface seasonal variation scale probably indicates a thickening trend in the corresponding active layer, which we will discuss in the chapter 6.
Figure 5.10: Seasonal variation scale in one full thawing season. For both periods. The negative sign represents increment in seasonal variation scale. The unit is mm and the color bars are in the same scale. (picture rotated 90° clockwise).
Figure 5.11: Statistic histogram of seasonal variation scale in both periods. The blue one is for period 1995-1999, the red one is for period 2006-2009. The unit is mm.
Figure 5.12: The difference in seasonal variation scale between the two periods. The negative sign represents the increment in surface seasonal variation scale. The unit is mm. (picture rotated 90° clockwise).
Figure 5.13: The long-term subsidence rate map over Zackenberg valley for both periods. The negative sign represents subsidence. The unit is mm/yr. The color bars are in the same scale.

5.4 Zackenberg surface deformation

To maximize the add-value of this study, it is valuable to extract deformation close to the location of ZERO programme. Therefore, the same analyzing strategies have been applied within the zackenberg valley. Figures 5.13(a), 5.13(b), 5.14, 5.15(a), 5.15(b), and 5.16 show the corresponding results in the zacekenberg valley.
Figure 5.14: Statistic histogram of subsidence rate over Zackenberg valley for both periods, the blue is for periods 1995-1999, the red is for periods 2006-2009. The unit is mm/yr.
Figure 5.15: The seasonal variation scale map over Zackenberg valley for both periods. The negative sign represents increment in surface seasonal variation scale. The unit is in mm. The color scale is in the same scale.

Over Zackenberg valley, during the period 1995–1999, the mean and standard deviation value of long-term subsidence rate is $-1.3 \pm 1.0$ mm/yr. In the period 2006–2009, this value increases to $-1.7 \pm 1.3$ mm/yr. In general, the magnitude of subsidence rate during 1995–1999 is about 0.3–2.4 mm/yr, and it rises up a little to 0.4–3.1 mm/yr during the period 2006–2009.

In the period of 1995–1999, the mean and standard deviation value of seasonal variation scale is $-3.3 \pm 2.9$ mm. During 2006–2009, the value increased to $-5.9 \pm 4.6$ mm. The statistics indicate the magnitude of seasonal variation is about 0.3–6.2 mm during 1995–1999, and the magnitude increases to 1.3–10.6 mm during 2006–2009.
Figure 5.16: Statistic histogram of seasonal variation scale over Zackenberg valley for both periods. The blue is for periods 1995–1999, the red is for periods 2006–2009. The unit is mm.
Chapter 6

PERMAFROST CHANGES AND METHANE FLUXES

In this chapter, based on the InSAR observed surface deformation, we have first estimated a general magnitude of permafrost thawing rate. We have also estimated the ALT when it reaches maximum development within the thawing season. Then we calculate both the total area where permafrost is under a thawing pattern and the total thawed permafrost in terms of volume loss during the corresponding periods. Assuming a linear relationship between the thawed permafrost and released methane flux observed in the same area, we have estimated the magnitude of total released methane flux over the whole study area in a certain period.

6.1 Permafrost changes

Permafrost thawing rate Permafrost thawing causes the long-term surface subsidence. This is corresponding to a volume contraction from ice to water. The ground ice content at this region is about 5–15% (percentage by volume) [9]. Since we already have the estimations on the long-term surface subsidence rate, the corresponding permafrost thawing rate can be estimated using equation 6.1.

\[
R_{thaw}^\ast = \frac{\hat{R}}{P_{ice} \cdot F_{vol}}
\]  

(6.1)

Where \( R_{thaw}^\ast \) indicates the estimated permafrost thawing rate. \( \hat{R} \) is the estimated surface long-term subsidence rate. \( P_{ice} \) is the ground ice-content. \( F_{vol} \) is the volume contraction factor (from ice to water). Using this equation, we found a general magnitude of permafrost thawing rate of 17.0±8.4 cm/yr during 1995-1999 and a thawing rate of 24.0±12.0 cm/yr during 2006-2009.
We already know all founded thawing locations together with their estimated thawing rate. The pixel size after resampling is 100x100 m². We then calculate the magnitude of total thawing area. Further, given a fixed period, the total thawed permafrost (in volume) can be estimated afterwards. Table 6.1 shows the estimated results.

<table>
<thead>
<tr>
<th>Period</th>
<th>$S_{thaw}$</th>
<th>$S_{InSAR}$</th>
<th>$S_{study}$</th>
<th>$R_{thaw}$</th>
<th>$V_{loss}$</th>
</tr>
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<tr>
<td>1995-1999</td>
<td>506.1</td>
<td>1074.4</td>
<td>5041</td>
<td>17.0 ± 8.4</td>
<td>2.1 · 10^8 ± 1.1 · 10^8</td>
</tr>
<tr>
<td>2006-2009</td>
<td>633.9</td>
<td>1392.2</td>
<td>4900</td>
<td>24.0 ± 12.0</td>
<td>3.0 · 10^8 ± 1.5 · 10^8</td>
</tr>
</tbody>
</table>

Table 6.1: A table of estimated thawed permafrost in volume during the corresponding periods.

Where $S_{thaw}$ represents the area of thawing. $S_{InSAR}$ is the area where we have InSAR observations. $S_{study}$ indicates the total size of study area. $R_{thaw}$ is the estimated permafrost thawing rate. $V_{loss}$ is the estimated permafrost loss (in volume) during the corresponding periods.

Active layer thickening The increment in surface seasonal variation scale indicates a thickening trend in the ALT. We have already estimated the surface seasonal variation scale ($\hat{S}_v$). We use the same $P_{Ice}$ and $F_{vol}$ as used in equation 6.1. Then, the estimated ALT is obtained using equation 6.2.

$$\hat{ALT} = \frac{\hat{S}_v}{P_{Ice} \cdot F_{vol}}$$  \hspace{1cm} (6.2)

Where $\hat{ALT}$ is the estimated active layer thickness. $\hat{S}_v$ is the surface seasonal variation scale. The estimated ALT over the whole study area in 1995-1999 is about 42.5 ± 26.3 cm while 68.0 ± 34.1 cm in 2005-2009. Based on our estimations, we found the ALT had thicken at a average magnitude of 25.5 cm from the period 1995-1999 to the period 2006-2009 over the study area.

Zackenberg ALT measurements ZERO programme has in-situ measurements on the ALT with respective to the surface at two different places. One of their measurements is shown in figure 6.1. The measurement is carried out at location [74°47’N, 20°56’W]. The grid is about 100 m by 100 m wide. The thickness of active layer has been measured each year during the thawing season since 1997 (From the beginning of June to the end of August). Usually, the active layer thickness reaches its maximum development at the end of August in each year. Based on the measurement during the period 1997–2009, the active layer thickness is increasing over years. According to their measurements, in the period 1997-1999, the average ALT when
it reaches maximum development near the end of thawing season is about 62 cm, while in period 2006-2009, the averaged maximum ALT increases up to 78 cm.

We take our estimations of surface seasonal variation scale at the same location as the measurement is taken. Figures 6.2(a) and 6.2(b) show the estimated value for both periods.

The location of the upper estimation is taken at [74°46.5N, 20°57.8W], the estimated seasonal variation scale is about 6 mm in the period 1995-1999. The location of the lower one is taken at [74°46.5N, 20°57.8W], the estimated seasonal variation scale is about 8 mm in the period of 2006-2009. Based on our estimated values, using equation 6.2, we found the estimated ALT in period 1995-1999 is about 40–60 cm; while the estimated ALT in period 2006–2009 is about 53–80 cm.

In general, comparing our estimations on ALT at the same location with their in-situ measurements, they both agrees with each other in the same order of magnitude.

Figure 6.1: ZEROCALM-1 in-situ measurements of ALT, taken at location [74°47N, 20°56W] (measurements taken by ZERO).
Figure 6.2: The estimated time-series profiles at the same location of ALT measurement. The upper one (a) is in 1995–1999: the estimated seasonal variation scale is about 6 mm. The lower one (b) is in 2006–2009: the estimated seasonal variation scale is about 8 mm.
6.2 Fluxes estimation

In the year of 1997, Christensen [17] has carried out a gas flux monitoring programme within the Zackenberg valley. The gas flux measurement is situated at the location [74°50'N, 20°54'W]. This area is of tundra wetland, which is about 270 x 360 m² wide, including five types of vegetation [17]. The measurement has been carried out from 19-Jun to 18-Aug in 1997 (about 60 days). Their computation revealed a methane source of \(1.9 \pm 0.7 \text{ mg} \cdot \text{m}^{-2} \cdot \text{hr}^{-1}\), which is the integrated mean daytime flux values across all vegetation types of the entire valley [17].

Since methanogenesis is highly sensitive to temperature, methane fluxes from tundra wetlands are expected to increase with temperature rise [34] and with the release of methane from thawing permafrost [80]. Although thickening of ALT is also expected to release methane flux, here we assume the observed methane flux is mainly due to thawed permafrost.

Researchers have reported a strong linear [69], curvilinear [6] or log-linear [20, 35] relationship between methane emission and soil temperature for tundra wetlands. Halsey [31] have found a linear relationship between the mean annual temperature and the permafrost degradation. Figure 6.3 plot the mean annual temperature against the percentage permafrost coverage.

![Figure 6.3: The estimated linear relationship between the annual mean temperature and the percentage permafrost cover [31].](image)

Therefore, based on the linear relation between methane emission and soil temperature as well as the linear relation between temperature and permafrost degradation. We have made an overall assumption for a linear relationship between released methane flux and permafrost thawing rate. This linear relationship is expressed in equation 6.3.
\[ \hat{k} = \frac{R_m}{\hat{R}_{thaw}} \]  \hspace{1cm} (6.3)

Where \( k \) is the estimated factor which indicates the linear relationship between these two observations. \( R_m \) is the observed methane releasing rate. \( \hat{R}_{thaw} \) is the estimated permafrost thawing rate. We consider an area centroid at the location of the gas flux measurement with a radius of 2 km to cover the main part of Zackenberg valley. We take the mean value of estimated permafrost thawing rate for all pixels within this area as our integrated mean value across the entire valley. The area and the location of gas measurement are indicated in figure 6.4.

![Location of Gas flux measurement within Zackenberg valley](image)

Figure 6.4: Location of Gas flux measurement within Zackenberg valley, indicated by the gray square box. The area is indicated by the black shaded area. They are superimposed on the permafrost thawing rate measurements.

The calculated mean value of estimated permafrost thawing rate in this area is 7.6 cm/yr. the observed methane flux rate is reported as 1.9 ± 0.7 mg · m⁻² · hr⁻¹. Then the relationship factor \( k \) can be obtained using equation 6.3. Since the gas flux measurement is only from 19 – Jun to 18 – Aug in 1997, the computed mean gas flux is only through daytime.
Thus only 12 hours is as an effective methane emission period in one single day. The total magnitude of released methane during that specific period can be obtained using equation 6.4.

\[
M_{meth} = \sum_{i=1}^{n} k \cdot \hat{R}_{thaw,i} \cdot S_i \cdot T
\]  

(6.4)

Where \(M_{meth}\) is the total mass of released methane flux. \(k\) is the relationship factor. \(R_{thaw,i}\) is the estimated permafrost thawing rate at \(i^{th}\) pixel (or called location) where we found permafrost thawing during 1995-1999. \(S_i\) is the area of \(i^{th}\) pixel (100x100 m\(^2\) for each pixel after resampled). \(n\) is the number of pixels. \(T\) is the same period as the methane flux had been observed (12 hours in one single day, about 60 days in the year 1997). Our estimation on total methane fluxes gives the magnitude of 1081.7 ± 398.5 T. This is the estimated magnitude of methane fluxes during the period 19-Jun to 18-Aug in 1997 over the whole study area. The value may be significant but give an overview of the total released methane in certain area during specific period.
Chapter 7

CONCLUSION AND FUTURE WORK

In this chapter, we give a summary on all the results that have been found during this study. The research question has been answered, followed by some recommendations for future research.

7.1 Conclusion

The research question in this study is to investigate the changes in permafrost as well as its magnitude. We selected an area of interest covering Zackenberg NE Greenland. First, the MT-InSAR technique was applied to obtain the time-series surface deformation over the last two periods: 1995-1999 and 2006-2009. Then we used a deformation model to fit our observations. We extracted two kind of deformation signals from this modelling process. One is the long-term surface subsidence rate, which infers thawing in permafrost layer; another is the surface seasonal displacement, which infers the variation in the active layer thickness. Given the ground-ice content over the study area, we have estimated the magnitude of the corresponding permafrost thawing rate and the average active layer thicknesses over the study area during the last two periods respectively. Our estimation on ALT at Zackenberg valley has agreed with their in-situ measurement at the same place. Further, assuming a linear relation between thawed permafrost and observed methane flux in Zackenberg valley, we have also given an estimation on the total scale of released methane over the whole study area during Jun.-Aug. in the year of 1997.

In summary, over the whole study area (around 70 x 70 km$^2$), about one fourth is covered by InSAR measurements. With all the observations and estimations, the research question is answered in the following aspects:
Surface deformation  Over the whole study area, during 1995–1999, we found a long-term surface subsidence rate of $1.3 \pm 1.0$ mm/yr and a surface seasonal displacement of $3.2 \pm 2.8$ mm. During 2006–2009, we found the long-term surface subsidence rate goes up a little to $1.8 \pm 0.9$ mm/yr. And the surface seasonal displacement increases to $5.1 \pm 2.8$ mm.

Permafrost change  Based on our estimation of surface deformation, correspondingly, we found a permafrost thawing rate of $17.0 \pm 8.4$ cm/yr and the ALT of $42.5 \pm 26.3$ cm during 1996–1999. About 506.1 km² area was under thawing and $2.1\cdot10^8 \pm 1.1\cdot10^8$ m³ permafrost had thawed. While in period 2006–2009, we found a permafrost thawing rate of $24.0 \pm 12.0$ cm/yr and the ALT of $68.0 \pm 34.1$ cm. 633.9 km² area was under thawing and $3.10^8\pm1.5\cdot10^8$ m³ permafrost had thawed. Comparing these two periods, we found an acceleration in permafrost thawing rate at the order of $0.7$ cm/yr². We also found a thickening trend in the active layer. This trend has been observed by Zackenberg in-situ measurement, and our estimation on ALT agrees with their in-situ measurement.

Methane flux  We also included an estimation on the total magnitude of methane flux released from the whole study area within a specific period. Assuming a linear relation between thawed permafrost and observed methane flux, the estimated magnitude of methane flux is of $1081.7 \pm 398.5$ T during Jun.–Aug. in 1997.

7.2 Future work

Since in the study area, only one fourth of pixels is selected by small baseline approach. In terms of improving the spatial density of MT-InSAR observation, a recommend action would be the idea similar with partial-PS [36].

Since we do not have much SAR acquisitions within thawing season to be used in deformation modelling, the estimation might be biased if outlier exists. In terms of improving the reliability of surface deformation estimations, we recommend to include more SAR observations covering larger time scale to minimize individual outlier contribution as more SAR data (especially L-band data) from various satellites become available. In terms of improving the model, we recommend to use a temperature-based thawing season period rather than a fixed one as well as a flexible seasonal varying scale in each thawing season during deformation modeling. We also recommend to check the soil moisture records, since variation in soil moisture could lead to changes in radar penetration depth, this changes lead to false abrupt subsidence or uplift signals in InSAR observations, thus it is recommended to correct this before modeling.
Since we assume the same linear relation between thawed permafrost and released methane flux over the whole study area, this might be not strong enough because of other process, such as root respiration in active layer, also release methane. Moreover these processes varies a lot in both spatial and temporal domain. In terms of improving the reliability of methane fluxes estimation, we recommend to consider the correlations with other important environmental factors, such as soil temperature & moisture, vegetation types, water table and such. It is also recommended to use experimental model to include the spatial variation of methane flux on the basis of different environmental settings.

In terms of validation of our deformation estimations over the whole area. We recommend to involve other geodetic method to set up a continuous displacement monitoring within the reference area as a ground truth point for InSAR measurements, such as GPS. In case the reference point moves, we could easily correct our relative deformation measurement to that amount of displacement. We also expect to have more site-measurement data available for us to validate our estimations on permafrost change and methane fluxes.

Above all, for improvements, there are several aspects need to be done in the near future, any tiny further step ask for more researches and efforts.
Bibliography


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APPENDIX A: ALT site measurements (ZERO)
APPENDIX B: Soil temperature records (ZERO)

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