Spatial and temporal variations of
air temperature inversions over different surface types
on Ammassalik Island (East Greenland)

Master Thesis

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ABSTRACT

Air temperature inversions, situations in which atmospheric temperature increases with height, are a common feature in the Arctic planetary boundary layer. Such stable layers have multiple consequences for the Arctic environment. Besides affecting vertical gradients of flora and fauna, they also have a direct impact on physical characteristics such as permafrost thaw depths, snow cover, and ice ablation. Comprehensive knowledge of the spatial and temporal variability of temperature inversion characteristics such as thickness, height, strength, and frequency are therefore crucial for understanding the surface impact of Arctic climate change.

The present study investigates the spatial and temporal variations of temperature inversions over different surface types and during different synoptic scale events on Ammassalik Island in East Greenland. During a field campaign in summer 2019, high temporal resolution profiles of atmospheric variables such as air temperature, humidity, and pressure were collected using unmanned aerial vehicles (UAVs). 113 profiles were acquired in a period of 13 days (06.07.2019-18.07.2019) over different surface types (rock, gravel, snow, ice) and at varying distances from the ocean (approximately 0km-6km).

Distinct air temperature inversions were present in 83% of the profiles, showing considerable variation in height and thickness. The rather bare (rocky and sandy) glacier foreland between the sea and the glacier induces surface heating and convection during the day and, through interaction with sea breezes, leads to the frequent formation of elevated inversions (EI). In contrast, the glacier surface itself (snow and ice) acts as a persistent cooling surface and leads to the formation of permanent surface-based inversions (SBI). The EI were most frequent and intense at times when changing air masses interacted with each other and the sea breeze was moderate. On cloudy and windy days, EI had lower frequency and intensity. The high-frequency and high-resolution profiles collected were compared with profiles from the nearby Tasiilaq radiosonde and with reanalysis data from ERA-I and ERA-5. All data sets (radiosonde and reanalysis products) show cold or warm temperature biases in the near-surface regions, which strongly affect the inversion statistics. The data on the ablation of the nearby Mittivakkat glacier were collected during the field campaign in order to gain a better understanding of the impact of temperature inversion statistics on ablation. It is concluded that incident solar radiation was one of the main drivers of ablation during the field campaign. A relation between strength, height and thickness of the EI and strength and thickness of the SBI on the glacier and glacier melt was observed.

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TABLE OF CONTENTS

	Page
List of Tables	v
List of Figures	vi
1 Introduction	1
1.1 Background	1
1.2 Theoretical foundations	3
1.2.1 Surface radiative cooling	3
1.2.2 Subsidence	4
1.2.3 Advection (warm and cold fronts)	4
1.2.4 Types of arctic inversion	4
1.3 Structure of the thesis	5
2 Methodology and Data	6
2.1 Research site	6
2.2 Meteorological observation	9
2.3 Atmospheric profiling	11
2.3.1 Unmanned aerial vehicles	11
2.3.2 Radiosonde measurements	17
2.3.3 Reanalysis data	19
2.4 Additional data	21
2.5 Data processing	23
2.6 Temperature inversion detection algorithm	29
3 Results and Discussion	31
3.1 Temperature characteristics of different data sets at near-surface levels . . .	31
3.1.1 Vertical temperature profiles	32
3.1.1.1 Radiosonde	32
3.1.1.2 ERA5 and ERA-I Model Level	35
3.1.2 Inversion characteristics	39
3.1.2.1 Radiosonde	39
3.1.2.2 ERA5 and ERA-I Model Level	41
3.2 Impact of surface types and settings on temperature inversions	44
3.2.1 Vertical profiles	44
3.2.2 Profiles along the surface	52

3.3	Impact of synoptic scale events on temperature inversions	55
3.3.1	Classified synoptic scale events	55
3.3.1.1	08.07.2019: Occlusion with cold front character	55
3.3.1.2	09-11.07.2019: Cold sector of a depression	57
3.3.1.3	13.07.2019: Ridge	58
3.3.1.4	15.07.2019: Trough	58
3.3.1.5	16.07.2019: Warm sector of a depression	59
3.3.2	Effect of synoptic scale events on inversion activity and intensity . . .	60
3.4	Impact of temperature inversion statistic on ablation	67
3.4.1	Observed ablation variability	67
3.4.2	Interrelations	72
4	Conclusions and Outlook	75
	Bibliography	78

LIST OF TABLES

TABLE	Page
2.1 AWS specifications	10
2.2 iMet sensor specifications	12
2.3 iMet GPS specifications	13
2.4 Temperature deviations between the sensors	13
2.5 Specifications of the locations A to H.	15
2.6 Dates of simultaneous UAV and RS measurements	16
2.7 RS station specifications	18
2.8 Main differences between ERA5 and ERA-I (ECMWF 2020d).	20
2.9 Pressure sensor and measurement specifications (LEIDERDORP 2020)	21
3.1 RMSE and the correlation coefficient between UAV and RS data	32
3.2 RMSE and the correlation coefficient per 100m altitude class	33
3.3 RMSE and the correlation coefficient between UAV and corrected RS data	34
3.4 RMSE and the correlation coefficient between the UAV and reanalysis products	36
3.5 RMSE and the correlation coefficient between the UAV and reanalysis products per 100m altitude class	36
3.6 Inversion statistic for UAV and RS data.	39
3.7 Inversion statistic for ERA-I and ERA5 data	41
3.8 Observed frequency of temperature inversions and inversion type in the UAV data	44

LIST OF FIGURES

FIGURE	Page
1.1 The different inversion processes	4
2.1 Location of the study area.	8
2.2 AWS used in the study	9
2.3 User interface of the DJI Pilot App	11
2.4 iMet-XQ2 sensor	12
2.5 Comparison of UAV temperature profiles	13
2.6 Position of the iMet-XQ2 sensor on the UAV	14
2.7 Surface types of the different locations A to H.	15
2.8 Pictorial overview of the different locations A to H	16
2.9 Location of the RS in Tasiilaq.	17
2.10 Illustration of the geopotential height at full model levels of ERA5	19
2.11 Automatically logging pressure sensor, measurement location and methodology at the proglacial lake	22
2.12 Measurement location and methodology at the proglacial river	22
2.13 Ablation stakes and dates of measurements	23
2.14 Comparison of six original with averaged UAV profiles	23
2.15 Temperature profile in a stable and unstable atmosphere	24
2.16 100m height segments of the UAV profiles in the colour of the calculated LR	25
2.17 The individual steps of the interpolation	26
2.18 Temperature variation at the different AWS	27
2.19 UAV, RS, ERA5 and ERA-I profiles	28
2.20 Detected inversion characteristics	29
2.21 Change in inversion thickness depending on thickness of included non-inversion layer	30
3.1 UAV profiles and RS profiles on one plot	32
3.2 Correlation between the RS and UAV profiles	33
3.3 Boxplot with deviation between UAV and RS	33
3.4 Correlation between temperature values of the RS surface level and of SYNOP	34
3.5 Correlation between the corrected RS and UAV profiles.	34
3.6 UAV profiles, RS profiles, ERA5 profiles on one plot	35
3.7 Correlation between ERA5, ERA-I and the UAV profiles	36
3.8 Deviation between UAV and reanalysis products per 100m altitude class	37

3.9	The first 50m of six original RS and corrected RS profiles	39
3.10	Inversion characteristics of UAV and RS data	40
3.11	The first 50m of six ERA-I and ERA5 profiles	41
3.12	Inversion characteristics of the UAV, ERA5 and ERA-I data	42
3.13	Observed frequency of temperature inversions and inversion type in the UAV data	44
3.14	Wind speed and direction of all 4 AWS	45
3.15	The temperature profiles of location A, B and C	46
3.16	Relationship between the intensity of the sea breeze and the inversion character- istics	47
3.17	The temperature profiles of location D and E	48
3.18	Deviation of profiles D and E from each other	49
3.19	The temperature profiles of location E, F and G	51
3.20	Horizontal temperature and relative humidity profiles	52
3.21	Horizontal temperature profiles along the path	53
3.22	Meteorological data of all AWS measured during the field study	56
3.23	Interpolated temperature profiles	57
3.24	Inversion intensity and frequency during synoptic scale events	61
3.25	UAV LR per day and weather class	62
3.26	Interpolated potential temperature profiles	63
3.27	Pearson correlation coefficient between WT, LL and meteorological variables . .	68
3.28	Hourly meteorological data from AWS GEUS	69
3.29	The measured ablation at ablation stakes 1 to 5	71
3.30	The approximate position of the glacier tongue in 1933	72
3.31	Daily inversion statistics	73

INTRODUCTION

1.1 Background

Air temperature inversion, a situation in which atmospheric temperature increases with height, is a common feature in the Arctic planetary boundary layer (KAHL 1990; MERNILD & LISTON 2010; SHAHI et al. 2020). Such inversions create stable layers of air that have multiple consequences for the Arctic environment. They impact vertical gradients of flora and fauna and have a direct effect on physical characteristics such as permafrost, thaw depths, and snow cover. Comprehensive knowledge of the spatial and temporal variability of temperature inversion characteristics such as thickness, intensity, strength, and frequency is therefore crucial for ice and snow melt calculations (CHUTKO & LAMOUREUX 2009), glacier mass-balance estimates (HULT et al. 2010; MERNILD & LISTON 2010), water resource predictions (ARCHER 2004), and pollutants concentration in mountain and basin areas (KERMINEN et al. 2007; KUKKONEN et al. 2005).

For this reason, many studies have focused on the quantification of inversion characteristics in the Arctic region (KAHL 1990; SEEREZE et al. 1992; MERNILD & LISTON 2010; ZHANG et al. 2011; SHAHI et al. 2020). Nevertheless, atmospheric vertical soundings have rarely been carried out in the Arctic due to the generally rough terrain, difficulty of access, and logistic constraints (MERNILD & LISTON 2010). Most of the studies mentioned above employed measurements from the few existing RS stations or modelled reanalysis data for their research.

Using RS data for the analysis of temperature inversions in the Arctic has certain advantages, such as long time-measurement series and high accuracy, but alongside the benefits of relatively inexpensive and continuous data collection, there are, however, several disadvantages. One is the low vertical resolution, as a result of which many of the shallower or near-surface inversions are not captured by the RS data (PALO et al. 2017; FOCHESSATTO 2015; MERNILD & LISTON 2010; VIHMA et al. 2011; SKONY et al. 1994). Furthermore,

RS data are not well suited to the analysis of diurnal or spatial changes in the atmosphere because of the low temporal (typically not more than twice a day) and spatial resolution of the data (MERNILD & LISTON 2010). Compared to RS data, modelled reanalysis data provide a higher horizontal and vertical resolution, represent a spatially and temporally coherent dataset, and cover long term periods, opening up new possibilities in Arctic temperature inversion research. However, despite all the advantages of the reanalysis data, there are several limitations. Previous studies show that reanalysis data may face problems in resolving lower troposphere conditions under extremely stable conditions (PALARZ et al. 2020; TJERNSTRÖM & GRAVERSEN 2009; MEDEIROS et al. 2011). Moreover, several cold or warm biases were detected in the reanalysis data, which can lead to incorrect quantification of temperature inversion characteristics (PALARZ et al. 2020; TETZNER et al. 2019; WESSLEN et al. 2014; GRAHAM et al. 2019).

Further improvement in our understanding of these limitations and how they affect the quantification of temperature inversions is needed. To address this need, during a summer 2019 field campaign in Sermilik on Ammassalik island, we employed UAV as a new and promising method of making in-situ atmospheric research measurements (KRAL et al. 2018; VILLA et al. 2016; ELSTON et al. 2015), allowing us to collect high resolution spatial and temporal profiles of atmospheric variables such as air temperature, humidity, and pressure. One aim of the present study was to use these high-resolution UAV profiles to evaluate how accurately RS data from the nearby station in Tasiilaq and two reanalysis products (ERA5 and ERA-Interim) reproduce temperature inversions and temperature profiles.

Compared to other locations in Greenland, scientific investigations have a long history in Sermilik, starting in 1933 (HASHOLT et al. 2016). Several studies conducted in this area have addressed and quantified temperature inversions (MERNILD & LISTON 2010; MERNILD et al. 2006; MERNILD et al. 2008; HASHOLT et al. 2016) because of the importance of this data for many different research fields. However, to analyse temperature inversions most of these studies used RS data, which is subject to the previously mentioned limitations. In order to provide a more comprehensive picture of temperature inversions and their spatial and temporal variability around Sermilik, we obtained 113 UAV vertical and 12 horizontal atmospheric profiles in a period of 13 days (06.07.2019-18.07.2019) over different surface types (rock, gravel, snow, ice) and at varying distances from the ocean (approx. 0-6km). The second objective of the present study was to use the profiles obtained to evaluate the spatial and temporal pattern of temperature inversions and the influence of surface type and geographical setting on the temperature inversion characteristics.

The causal relationship between temperature inversions and prevailing weather conditions has been demonstrated by different studies (MILIONS & DAVIS 2008; KASSOMENOS & KOLETIS 2005; KUKKONEN et al. 2005; MARSHALL et al. 2007; WHITEMAN et al. 1998). None of those studies used data which had such a high degree of temporal and spatial resolution as the data we collected during the field campaign in July 2019. Therefore, the third objective of this work is to quantify the influence of four synoptic scale events observed

during the field campaign on the temperature inversion activity (fraction of occurrence) and intensity (thickness and strength). For this purpose, the previously quantified temperature inversion characteristics from the 113 acquired UAV profiles are used.

Temperature inversions can have complex effects on the surface energy budget of Arctic glaciers, depending on the glaciers' characteristics. A number of studies have observed higher ablation rates in the presence of temperature inversions (GARDNER et al. 2009; CHUTKO & LAMOUREUX 2009; TJERNSTRÖM et al. 2015), but other studies have detected a positive effect of temperature inversions on glacier melt (less melt) (BRAUN et al. 2004; MERNILD & LISTON 2010; HULTH et al. 2010). Challenges, therefore, remain in our understanding of the complex interrelation between temperature inversions and glacier ablation. Hence, another goal of this study is to improve our understanding of the effect of temperature inversions on snow and glacier melt. The Mittivakkat glacier, one of Greenland's most intensively studied glaciers, is located in the study region. The ablation of the Mittivakkat glacier during the field campaign was quantified using a network of ablation stakes on the glacier and water level data from a lake and river in front of the glacier. These data will be associated with the quantified inversion characteristics derived from the 113 atmospheric UAV profiles, to achieve the fourth objective of this work.

1.2 Theoretical foundations

Temperature inversion - a reversal (lat. *inversio*) of the normal temperature pattern in the troposphere, in which temperature decreases with height - occurs when a layer of warm air lies above a layer of cooler air. Such temperature inversions can develop under different circumstances (PIDWIRNY 2020; BAUMBACH 1990). The most common formation mechanism can be categorized as surface radiative cooling, subsidence, and advection. This causes two different types of inversion that play a prominent role in the Arctic: inversions with an elevated base (EI) and inversions with a base at ground level (SBI).

1.2.1 Surface radiative cooling

Daytime solar radiation heats the land surface of the Earth during clear and sunny days. As the surface temperature increases, energy is conducted from this warmer surface to the cooler air above this surface and into the cooler soil below. Over the course of the day, more and more solar energy is absorbed by the surface and the temperature both at and above the surface increases. After sunset, solar energy no longer heats up the surface. The surface temperature starts to drop, due to long-wave radiation loss from the surface to the atmosphere. Under such conditions, the surface emits more energy than it receives from energy conduction from deeper soil or from atmospheric radiation. This process is especially strong in cloudless conditions due to reduced atmospheric radiation. Consequently, energy conduction from the warmer air above the surface to the surface takes place. Soon the air

adjacent to the surface becomes cooler than the air above (temperature increasing with height) (Figure 1.1 a) (HUPFER & KUTTLER 2006).

1.2.2 Subsidence

Air moves downward in high pressure areas and moves outward on reaching the ground. The subsiding (sinking) air warms adiabatically (compression) and dries during the descent. The top of the subsiding layer warms more than the lower part because it descends further through the atmosphere (AHRENS & HENSON 2016). This produces a situation in which the air temperature increases with increasing height (Figure 1.1 b). This temperature inversion mechanism tends to occur when there are anticyclonic conditions or high-pressure ridges (GILSON et al. 2018; SERREZE et al. 1992).

Figure 1.1: The different inversion processes: blue shading indicates colder air masses and red shading warmer air masses. a) the black arrows represent the infrared radiation from the surface into the atmosphere, while the red arrows show the energy conduction from the air to the surface. b) the blue arrows show the subsiding air. c) the red arrows represent warm air sliding over colder air during a warm front. d) the blue arrows represent the cold air sliding under the warm air during a cold front.

1.2.3 Advection (warm and cold fronts)

When cold air masses collide with warmer air masses due to a cold front, the colder air layers move under the warmer air layers on account of their higher density. This lifts the warmer air layers and creates an inversion (Figure 1.1 d). If warm air masses meet colder air masses due to a warm front, the warmer air layer pushes over the colder one and also forms an inversion (Figure 1.1 c) (MONKHOUSE 1970).

1.2.4 Types of arctic inversion

Surface-based inversions (SBI) (Figure 1.1a) form as a result of surface radiative cooling as described in 1.2.1. This inversion formation mechanism occurs frequently during the Arctic winter and also at night when solar radiation is zero or negligible (GILSON et al. 2018; DEVASTHALE et al. 2011; MARSHALL et al. 2007). Furthermore, SBI usually occur above

melting snow and ice (KAHL et al. 1992; MILLER et al. 2013; GILSON et al. 2018; SHAHI et al. 2020). The temperature above melting snow or ice is prevented from rising and therefore is cooler than the surrounding heated air masses (AYALA et al. 2015). Consequently, the air temperature rises with the increasing distance from the melting surface (if the surrounding temperature is higher than 0°C) and creates almost permanent SBI above snow and ice.

Elevated inversions (EI) (Figure 1.1 b, c, d) are formed by two processes: subsidence (1.2.2.) and advection (1.2.3.) (SHAHI et al. 2020; GILSON et al. 2018). EI are most common above the Arctic Ocean and around coastal areas during the Arctic melt season, but can occur at any time of the year (CHUTKO & LAMOUREUX 2009).

Both types of inversion, SBI and EI, can have complex effects on the atmospheric processes close to the ground and in the lower troposphere (SHAHI et al. 2020). For example, both are important for understanding fog or low cloud formation (GILSON et al. 2018; SED-LAR & TJERNSTRÖM 2009), the trapping and concentration of pollutants (SCHMALE et al. 2018; BRIDGMAN et al. 1989), surface destruction (BARRIE et al. 1988), the distribution of ozone (WESSEL et al. 1998), and snow and ice melt processes in the Arctic (MERNILD & LISTON 2010; CHUTKO & LAMOUREUX 2009; HULT et al. 2010). EI are also important for estimating the geostrophic drag coefficient (OVERLAND & DAVIDSON 1992), which is a parameter required for sea ice models (SHAHI et al. 2020 after HIBLER & BRYAN 1987). Studying the changes, properties, and variability of temperature inversions is thus crucial for understanding many atmospheric processes in the Arctic.

1.3 Structure of the thesis

The thesis is divided into four chapters. Chapter One (1) introduces the topic and summarizes the scientific and thematic background on which the thesis is based. Chapter Two (2) explains the data sources, the measurement site and measurement setup, and also the pre-processing of the data and the inversion detection algorithm employed. Chapter Three (3) describes and presents the research results and also includes the discussion of the results. The conclusion and suggestions for further work are contained in Chapter Four (4).

METHODOLOGY AND DATA

This chapter presents the data collected and used in this study and the methodology employed. First the research site is introduced (2.1). This is followed by a description of the data used in the study and how the data were collected (2.2, 2.3, 2.4) and processed (2.5). The chapter closes with a description of the temperature inversion detection algorithm which was applied (2.6).

Unless otherwise stated, all analyses, calculations, and graphs in this study were carried out using the programming language R and the graphical user interface or development environment RStudio. Furthermore, if not otherwise indicated, all images used in this thesis were created by Jakob Abermann or Iris Hansche.

2.1 Research site

Ammassalik Island is located in the Sermersooq municipality on the south-east coast of Greenland, centred at 65°N, 37°W (Figure 2.1). To the west, the 678km² island is separated from the mainland by the 10-15km wide Sermilik Fjord, which extends in a north-south direction and is the largest fjord system in SE Greenland; to the east it is bounded by the Ammassalik Fjord and to the north by the Ikaasartivaq Fjord; to the south it fronts the North Atlantic Ocean (CHRISTIANSEN et al. 1999; MERNILD et al. 2008; MERNILD et al. 2010). The primary research site for this study, the Mittivakkat area, is situated on the southwestern part of the island about 50km east from the eastern margin of the Greenland Ice Sheet (GrIS) and approximately 15km north-east of the town of Tasiilaq (formerly known as Ammassalik) (MERNILD et al. 2008; MERNILD & HASHOLT 2006). This area is characterized by sporadic permafrost and a strong alpine relief with elevation varying between 0m and 1096m above sea level (Vegas Fjeld) (CHRISTIANSEN et al. 1999).

The climate in the region where the study was conducted is strongly influenced by the passage of lows, the proximity of the Icelandic Low, and of the warm Irminger Current (GILSON 2018). The climate is also marked by a seasonal sea ice cycle and a strong annual

solar cycle. The midnight sun line crosses the Mittivakkat area, while the polar night line is approximately 200km further north (CAPPELEN et al. 2015; GILSON 2018). In addition to the influence of the macroscale weather phenomena, there are several local factors such as topography, fjord configuration, and distance from the coast which modify the climate in the study region (affecting, for example, the wind field). One important and frequent mesoscale thermal circulation in the region is governed by local temperature differences between land and sea. During the day, the land heats faster than the sea. Consequently the air above the land becomes warmer and produces a thermal low. The air above the sea remains cold and produces a thermal high. This pressure distribution pattern creates a sea breeze near the surface that blows from sea to land and, at higher altitudes, a compensating flow from land to sea (AHRENS & HENSON 2016). At night, the situation is reversed.

According to the Köppen classification system, the climate in the study region is classified as a tundra climate (ET) (MERNILD & HASHOLT 2006), with a mean annual air temperature (MAT) of about -1°C . Based on the data from the automated weather station (AWS) in Tasiilaq over the period 1981-2010 (climatic normal period), July is the warmest month, showing a mean temperature of 7°C close to the coast, and February the coldest month, with a mean temperature of -7°C . The mean annual precipitation in the study region is around 800-1000mm, with lower precipitation during the summer. Approximately 80-90% of the precipitation falls as snow (JAKOBSEN et al. 2013).

The name of the area is derived from Mittivakkat, a local, temperate glacier (centred at $65^{\circ}420\text{ N}$, $37^{\circ}480\text{ W}$), which is one of Greenland's most studied glaciers (YDE et al. 2014). In 2011, the glacier was approximately 5.5km long, with an average thickness of 115m, covering an elevation range between 150m to 880m a.s.l. and an area of 26.2km^2 (according to MERNILD et al. 2017) (MERNILD et al. 2013; MERNILD et al. 2008). Since the end of the Little Ice Age (approximately 1900 AD), the glacier has been continuously retreating (MERNILD et al. 2013). The glacier catchment drains through a proglacial valley (Mittivakkat valley), which is approximately 2km long and runs E-W from the glacier (CHRISTIANSEN et al. 1999). The main valley and all the connected valleys are the result of pulsed erosion during multiple cycles of glacier expansion and retreat during the Quaternary (HASHOLT et al. 2008; MACHGUTH et al. 2016).

In 1933 the seventh Thule Expedition visited Ammassalik island, performing ablation measurements on the Mittivakkat glacier that are among the first glacier measurements in Greenland (MACHGUTH et al. 2016; HASHOLT et al. 2016). In 1970, the University of Copenhagen established a permanent research station on the shore of Sermilik Fjord. The Sermilik Research Station was built to provide a logistics base for future field work and monitoring programmes. The University of Copenhagen also set up a small hut on a nunatak (515m a.s.l.) in 1993. Today, the Sermilik station supports a year-round programme monitoring glacier mass balance, sediment transport and discharge in the Mittivakkat valley, basic climatology, local climate gradients, a weather station transect, and delta and coastal geomorphological development (INTERACT 2017).

Figure 2.1: Location of the study area with the research station (green dot), ablation stakes (yellow dots), weather stations (black triangles), locations where water level measurements were taken (blue dots), the path walked with the backpack sensor (10.07.2019) (red dashed line) and the locations where the UAV profiles were taken (red dots). Outline map of Greenland: EntchantedLearning; Basemap: Sentinel-2 2019; Data processing in QGIS.

2.2 Meteorological observation

Air temperature, air pressure, humidity, incoming shortwave radiation, wind speed, and wind direction from 5 AWS were used to analyse and verify the RS data, to assess the changes in the meteorological parameters, and to classify the surrounding weather conditions in the area of the field campaign. The measurement specifications and other important information for all the AWS are presented in Table 2.1.

Four of the meteorological stations are sited within the study area (Figure 2.1). Station Coast (COAST) was established in 1997 by the University of Copenhagen (MERNILD et al. 2008). The AWS is located on a small hill near the sea and the Sermilik Research Station (Figure 2.2 a) and faces the ocean.

Figure 2.2: The AWS in the study area (a, b, c, d, e) and outside the study area (f).

Station 196 (196), established in 2009 by the University of Copenhagen (UC), is located on a hill close to the research station (430m) (Figure 2.2 b). We downloaded the data for the period 9 August 2017 - 19 July 2019 directly from both AWS COAST and AWS 196.

An AWS sited near the equilibrium line altitude (ELA) on Mittivakkat and operated by GEUS was established in 2009 (Figure 2.2 c). The weather station is part of the “Programme

for Monitoring of the Greenland Ice Sheet” (PROMICE) network in Greenland, which is headed by the Geological Survey of Denmark and Greenland (GEUS) with the goal of assessing the changes in the mass of the GrIS (FAUSTO et al. 2020). We downloaded the data from the period 5 July 2019 - 17 July 2019 from the AWS GEUS.

AWS Name	Location	M a.s.l.	Operator	Parameters	Time interval for measurements
COAST	65°41'N, 37°55'W	25	UC	air temperature, humidity, wind speed, wind direction, precipitation, incoming and reflected (outgoing) short-wave radiation	10 min.
196	65°40'N 37°55'W	196	UC	air temperature, humidity, wind speed, wind direction, precipitation, incoming and reflected (outgoing) short-wave radiation, snow depth	10 min
GEUS	65°41'N 37°50'W	440	GEUS	air temperature, wind speed, pressure, humidity, downward and upward components of shortwave and longwave radiation, temperature in the upper 10m of the ice, station tilt, snow, ice surface height	10 min
NUNATAK	65°42'N 37°49'W	515	UC	air temperature, humidity, wind speed, wind direction, wind gust, precipitation, incoming and reflected (outgoing) short-wave radiation, snow depth	10 min
CCMORE	65°43'N 37°47'W	748	UG	air temperature, pressure, humidity, wind speed, wind direction, precipitation, dew point, precipitation type, wet bulb temperature	5 min
SYNOP	65°36'N 37° 38'W	44	DMI	air temperature, pressure, humidity, wind speed, wind direction, precipitation, incoming short-wave radiation, dry bulb temperature, cloud cover	1 hour

Table 2.1: AWS specifications.

One station in the area was not used for this study, Nunatak Station (Figure 2.2 d). The station is located on the highest point of a nunatak on the northern side of the Mittivakkat glacier and was close to the ELA at the time of installation (HASHOLT et al. 2004; MERNILD et al. 2008). However, the station stopped collecting data in January 2019 and the data collected before then showed significant data gaps. The station was visibly damaged and some sensors destroyed. Data from this station were therefore not used in this study.

In order to have a weather station transect (Figure 2.1) with stations at different elevations, we established AWS CC-MORE (CCMORE) on a nunatak at 748m a.s.l. (Figure 2.2 e). The CC-MORE weather station was set up on 7 July 2019 and collected data for 8 days.

Tasiilaq station (SYNOP) is a standard synoptic World Meteorological Organization (WMO) meteorological station operated by the Danish Meteorological Institute (DMI). The station is not located in the study area. It is located approximately 15km away on a hillside at the upper town boundary of the Tasiilaq settlement (Figure 2.1) (Figure 2.2 f) (MERNILD & LISTON 2010). The station was established before 1895. The data are quality controlled and distributed by the DMI (CAPPELEN et al. 2019). The air temperature data from Tasiilaq station were used to assess the quality of the RS surface level values.

2.3 Atmospheric profiling

2.3.1 Unmanned aerial vehicles

Originally, the development and application of Unmanned Aerial Vehicles (UAVs) was driven by the military, but during recent decades UAVs have been used in other fields such as rescue missions, petroleum and mineral exploration, ecological and environmental studies, agriculture, mapping, and meteorology (COOK et al. 2013). The recent development of small, lightweight, and low-cost UAVs and meteorological sensors in particular opens up new opportunities for meteorological measurements in the field of atmospheric research (KIMBALL et al. 2020; KRAL et al. 2018; ELSTON et al. 2015; VILLA et al. 2016; REUDER et al. 2012).

Figure 2.3: Screenshot of the user interface of the DJI Pilot App.

The ability of UAVs to measure with high temporal and spatial resolution, their flexibility in use and ability to fly in situations that are too difficult for manned aircraft (e.g., low heights above ground, during thunderstorms) make them the ideal choice for studying the Earth's atmospheric boundary layer (ABL) (KIMBALL et al. 2020). They have therefore been widely used to measure several ABL features such as aerosols (CORRIGANE et al. 2008; VILLA et al. 2016); thermodynamic quantities such as temperature or pressure (REUDER et al. 2012;

COOK et al. 2013); wind speed and direction (GARREAU 2020; HOUSTON & KEELER 2018) and heat flux (LEE et al. 2018).

The UAV model used in this study was the DJI MAVIC Pro multirotor aircraft. The significant advantages of multirotor UAVs over fixed-wing UAVs are their ability to hover in the air, to land and take off on a small spatial footprint, and their manoeuvrability (VARENTSOV et al. 2019). DJI MAVIC has 4 rotors, weighs 734g, has a maximum ascent speed of 5m/s, maximum descent speed of 3m/s, maximum flight speed of 64.8km/h, maximum flight time of 27 minutes, and has an integrated HD camera to record images and videos. The maximum flight altitude in this study was approximately 500m a.g.l.

The UAV was controlled by a remote controller connected to a smartphone. Both systems could communicate through the DJI Pilot app, which is fully compatible with any iOS and Android system. The app allows the pilot to navigate the UAV (Figure 2.3), to change flight settings, and to change the angles of the camera (DJI 2020). We used three DJI MAVIC UAVs in this study, two for performing the measurements and one as a backup.

In this study, pressure, temperature, and humidity up to an altitude of 500m a.g.l. were measured using an International Met System iMET-XQ2 sensor (Figure 2.4). The sensor specifications and GPS specifications are presented in Table 2.2 and Table 2.3 (INTERMET 2018). The data col-

Figure 2.4: *iMet-XQ2 sensor (INTERMET n.d.).*

lected were stored on an on-board 16Mb flash memory card and could be downloaded via a USB interface. In order to read and view the data collected iMet-X software must be installed. The data can then be exported and stored as a txt or csv file.

Sensor	Range/Resolution	Response Time	Accuracy
Pressure	1200hPa to 10hPa / 0.02hPa	< 1 second	± 1.5 hPa
Temperature	-95°C to 50°C / 0.01°C	< 1 second	± 0.3 °C
Humidity	0% to 100% / 0.1%	0.6 seconds	± 5 % RH

Table 2.2: *Sensor specifications (INTERMET 2018).*

Range/Resolution	Horizontal Accuracy	Vertical Accuracy	Update Rate
40km/0.01m	8m	12m	1 second

Table 2.3: *GPS specifications (INTERMET 2018).*

In all, we used 3 iMET sensors in this study. The humidity sensor of iMET sensor number 1 broke on 13 July 2019 and the temperature sensor of iMET sensor number 2 on 17 July 2019. We became aware of both of these problems in the evenings of the same days. The profiles acquired with the broken sensors were excluded from later evaluations.

We conducted an experiment to find out how the temperature measurements obtained by the sensors differed from each other. In the experiment (Figure 2.5a), all sensors were fixed on a wire 34m below the UAV. The speed of ascent was 4km/h between 0m and 100m a.g.l. and 6km/h from 100m a.g.l. to approx. 500m a.g.l.

Ascent			
Sensor	Mean (°C)	St. Dev. (°C)	N
Sensor 1 – Sensor 2	- 0.14	0.24	388
Sensor 1 – Sensor 3	0.00	0.19	517
Sensor 2 – Sensor 3	0.06	0.19	650
Descent			
Sensor 1 – Sensor 2	0.00	0.31	574
Sensor 1 – Sensor 3	- 0.04	0.32	763
Sensor 2 – Sensor 3	0.06	0.27	683

Table 2.4: *Temperature deviations between the sensors.*

During the descent, a speed of approximately 4km/h was maintained throughout. The average deviation between the sensors during the ascent and descent is shown in Table 2.4. It can be seen that the mean temperature difference between the sensors when measuring at the same heights is small. Figure 2.5b shows that the temperature differences are greatest in the area where the inversion begins. During the ascent, sensor 2 registers the temperature increases approximately 30m higher than sensors 1 and 3. During the descent, the temperature curves of the sensors show greater agreement, and the curve is also more homogeneous in the first 150m a.g.l.

Figure 2.5: *The temperature profiles during the experiment of all sensors (a) during ascent and descent and the temperature differences between the sensors (b) also during ascent and descent.*

Data acquisition and measurement setup

The field campaign on Ammassalik Island started on 6 July 2019 and lasted for 13 days until 18 July 2019. During this period we generated 113 profiles, with high temporal and spatial resolution, of atmospheric variables such as air temperature, humidity, and pressure above different measurement points at up to 500m a.g.l. The measurements were carried out using UAVs equipped with iMET-XQ2 sensors (Figure 2.6).

The studies conducted by KIMBALL et al. (2020) and GREENE et al. (2018) investigated the effect on sensor temperature measurements of mounting sensors in different positions on UAVs. Based on their results, the best position would be 5cm-10cm below the rotor arms, one third along the length of the propeller. Such a position provides enough aspiration (contact with surrounding air) while preventing contact with the warm air stream from the propeller tips and the motor (KIMBALL et al. 2020; GREENE et al. 2018). Such sensor positioning was not possible on the UAV DJI MAVIC Pro that was used. The sensor was not small enough to fit beneath the rotor arms without being hit by the propeller because the propeller of the UAV that was used was longer and the rotor arms shorter than the one used in the studies.

In order to ensure adequate ventilation, the iMet-XQ2 was placed on top of the UAV in between the rotors (Figure 2.6). In this position the sensor would be aspirated by the propeller wash. Furthermore, as suggested by INTERMET (2018), the iMet-XQ2 was oriented with the label-side pointing up. Since the components that create heat are located in the rear half of the UAV body, the sensor is mounted in the front half in order to avoid artificial sensor heating (KIMBALL et al. 2020; GREENE et al. 2018). Additionally, iMet-XQ2 has an aluminium coating in order to reduce the heating effect from terrestrial and solar radiation.

Figure 2.6: *Position of the iMet-XQ2 sensor on the UAV.*

In this measurement setup, the atmospheric profiles were acquired at specific measurement points. One aim of this study is to investigate the spatial and temporal variations or patterns of temperature inversions along a transect that crosses different ground surface types at varying distances from the Sermilik Research Station. The defined transect goes from this research station along the Mittivakkat valley and the top of the Mittivakkat glacier and ends at AWS Nunatak. The points where the atmospheric profiles were taken are shown in Table 2.5 and Figure 2.7.

The UAV ascents were standardized before the field study began. First the UAV was positioned and the sensor attached. The UAV was started shortly after the sensor obtained GPS reception. The ascent speed to 20m a.g.l. was approximately 2km/h, from 20m to 100m a.g.l. approximately 4km/h, and from 101m to 500m a.g.l. approximately 6km/h. Once the maximum flight height was reached, two photographs were taken: one of the ground directly beneath the UAV and a 360-degree panorama. After the ascent, weather observations and information about the flight (e.g., time of take-off and landing, max. flight height) were noted in the field book.

Name	Distance from Sermilik [m]	Height [m a.s.l.]	Location	Surface
A	0	8	Sermilik	Rock
B	410	5	Beginning of the valley	Sand
C	1110	8	Valley head	Sand
D	1550	150	Ridge	Sand
E	1910	150	Glacier tongue	Ice
F	2750	290	Ablation stake	Snow, ice
G	4260	440	GEUS	Snow, ice
H	5570	505	Nunatak	Sand, rock

Table 2.5: *Specifications of the locations A to H.*

Figure 2.7: *Surface types of the different locations A to H.*

In addition, we conducted simultaneous ascents to obtain information about the atmosphere at two different locations at the same time (Figure 2.8). The differences in atmospheric stratification were measured at both ends of the Mittivakkat valley (shown as location B and C in the Figure 2.8). The other simultaneous measurements were taken at locations D and E, where D was at 150m a.s.l. in front of the glacier over sand and E at the same altitude but on the glacier tongue over melting ice.

Figure 2.8: Pictorial overview of the different locations A to H. Data from the locations in the red rectangles collected simultaneously (B & C and D & E) (ABERMANN et al. 2019).

This study also compares the UAV data that was collected with the RS data from Tasiilaq. Ascents were made at Sermilik Research Station (location A) twice daily at the same time as the RS launches (0000 UTC and 1200 UTC). Over 12 days 21 ascents were performed (Figure 2.5).

In addition to the vertical atmospheric measurements obtained with the UAV, the horizontal atmosphere along the transect was also measured. An iMET sensor was mounted on top of a backpack at approximately 1.5m a.g.l. and carried along the transect. This measurement of the atmosphere was carried out in both directions, on the way up and the way down, until the battery ran low. This sensor was intended to provide information on ground-level changes in the atmosphere along the transect, e.g., the shift from valley to glacier atmosphere.

10:00 (AM)	22:00 (PM)
	07.07.2019
08.07.2019	08.07.2019
09.07.2019	09.07.2019
10.07.2019	10.07.2019
11.07.2019	11.07.2019
12.07.2019	12.07.2019
13.07.2019	13.07.2019
14.07.2019	14.07.2019
15.07.2019	
16.07.2019	16.07.2019
17.07.2019	
18.07.2019	18.07.2019

Table 2.6: Dates of simultaneous UAV ascents and RS at 10:00 (AM) and 22:00 (PM) [UTC- 2].

2.3.2 Radiosonde measurements

In general, the radiosonde (RS) is a small, battery-powered, lightweight instrument package carried below a latex helium- or hydrogen-filled balloon up to an altitude of 30km. The speed of ascent is approximately 5ms^{-1} and the balloon usually weighs between 0.3kg and 1.2kg (DABBEREDT et al. 2003). The instrument package includes various sensors for measuring atmospheric variables such as temperature, pressure, and humidity. In addition to the directly measured atmospheric variables, wind speed and direction can be inferred from the shift in the GPS position of the RS (AHRENS 2007). The meteorological variables are recorded at set levels of pressure and other significant thermodynamic variables to capture changes in humidity or temperature gradients between the levels (DURRE et al. 2006; GILSON 2018). The set pressure levels include those specified by the WMO in 1996 (1000, 925, 850, 700, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, 20, and 10hPa) (ISIOYE et al. 2016). Additionally, surface observations at or near the launch site are included (DURRE et al. 2006). These data are sent to the ground station via radio transmitter for data collection (BROCK & RICHARDSON 2001). Such soundings are widely used for operational weather forecasting, but they also provide valuable in-situ atmospheric data for climate research purposes. According to DIRKSEN et al. (2014) there are approximately 1,000 soundings daily around the globe.

Figure 2.9: *Location of the RS in Tasiilaq (LAURSEN 2015).*

In this study, we use RS data from the Integrated Global RS Archive (IGRA) (DURRE et al. 2006). The IGRA is a compilation of quality-controlled RS observations from a global network of more than 1,500 stations (ZHANG et al. 2011). In such datasets, format problems, climatological outliers, accidental duplication of levels, physically implausible values, internally, temporally, and vertically inconsistent values across soundings and levels are checked and removed or adapted (DURRE et al. 2006). These datasets are updated on a daily basis and can be downloaded from the IGRA homepage.

Location	Operator	WMO ID	M a.s.l.	Parameters
Tasiilaq (65°36'N, 37° 38'W)	DMI	4360	52	temperature, humidity, windspeed, pressure, height, dew point temperature

Table 2.7: *RS station specifications.*

The RS station Tasiilaq (Figure 2.9, Table 2.7) is located close to the WMO meteorological station SYNOP. Nowadays the twice daily soundings at 0000 UTC and 1200 UTC are launched automatically and collect data at altitudes from 52m up to 32000m-35000m a.s.l. The first RS launches were carried out in Tasiilaq in 1947. From 1947 to 2019 a number of different sonde types were used: VIZ, VAISALA RS18, RS21, RS80, RS92 and RS41 (IGRA historical data). Currently the Vaisala RS RS41 is used at the station in Tasiilaq. According to VAISALA (2020), the sonde measures to an accuracy of 0.3°C, 4% RH, and 1hPa in the lower 16km.

Set against all the advantages of the RS as a tool to study temperature inversions in the Arctic, it nevertheless has several limitations. To quantify and analyse the limitations of the RS in capturing complex atmospheric stratifications, the two soundings per day (0000 UTC and 1200 UTC) at the Tasiilaq station were compared with UAV soundings collected simultaneously at Sermilik and the data from the nearby AWS (SYNOP). Quality is assessed in three ways. First, the RS surface temperature values are compared with temperature data from the nearby AWS SYNOP collected at the same time (0000 UTC and 1200 UTC) and altitude (52m a.s.l.). Secondly, temperature values at common data points (elevations) (from UAV and RS profiles) are compared to reveal general and altitude-dependent deviations that might show systematic deviations that could affect the detection of temperature inversions. Third, the ability and limitations of the RS in reproducing the inversion characteristic of the atmosphere are evaluated by comparing the detected inversion statistics of both datasets (UAV and RS). Only soundings at levels below 500m a.g.l. were used for the evaluation.

2.3.3 Reanalysis data

The ERA climate reanalysis, from the European Centre for Medium-Range Weather Forecasts (ECMWF), combines modelled data with observational data from across the world to produce consistent time series for a range of climate variables. The data assimilation principle follows the same method used by numerical weather prediction centres, by which previous forecasts are combined with newly available observations to produce a new best estimate of the state of the atmosphere and the surface beneath it (TETZNER et al. 2019; ECMWF 2020a).

In this study, we evaluate the ability of two reanalyses by ECMWF to reproduce the lower atmospheric stratification. These are the ERA-Interim (ERA-I) and the newly released (2017) fifth generation ECMWF reanalysis, ERA5 (TETZNER et al. 2019; GRAHAM et al. 2019). The ERA-I reanalysis is available with 0.7-degree horizontal resolution (approximately 80km) and six-hour temporal resolution for the period January 1979 to August 2019. Altogether ERA-I has 60 vertical model levels ranging from surface level up to 0.1hPa (approx. 80km) (GRAHAM et al. 2019; DELLHASE et al. 2020). ERA5 replaced ERA-I and was announced by HERBACH & DEE (2016) in April 2016. Like ERA-I, the ERA5 reanalyses extend back to January 1979, but have a greater temporal resolution of one hour, a

much greater horizontal resolution of 0.25-degree (approx. 31km) and a vertical resolution of 137 model levels from the surface up to 0.1hPa (approx. 80km) (PALARZ et al. 2018; GRAHAM et al. 2019; DELLHASE et al. 2020). In both reanalyses, the vertical resolution is finest in the planetary boundary layer and coarser nearer the top of the model. The lower levels (troposphere) follow the earth's topography and the higher ones (upper stratosphere and lower mesosphere) surfaces of constant pressure (ECMWF 2020a) (Figure 2.10). Another improvement over ERA-I is the integrated forecasting system (IFS) Cy41r2, which uses a form of Gaussian grid (octahedral grid). This gives increased horizontal resolution and thus a better representation of the topography and values close to the ground (ECMWF 2020b). Further information about the IFS improvements can be obtained from HERBACH & DEE (2016). The main differences between ERA5 and ERA-I are summarized in Table 2.8.

From both ERA5 and ERA-I we extracted data for air temperature and specific humidity for all available model levels. The values from the nearest grid cell to the observations (Sermilik) were linearly interpolated to the coordinates of the Sermilik station. Only the model levels beneath 500m a.g.l. from ERA5 and ERA-I were used for the evaluation (ERA-I: 10, ERA5: 16).

Figure 2.10: Illustration of the geopotential height at ERA5 full model levels from 1 to 91. The topography is represented by the grey area (HORAK 2019).

	ERA-I	ERA5
Period covered	1979 - 2019	1950 - present
Production Period	August 2006 – 31st August 2019	Jan 2016 – end 2017, then continued in near real-time
Assimilation system	IFS Cycle 31r2 4D-Var	IFS Cycle 41r2 4D-Var
Spatial resolution	79 km globally, 60 levels to 0.1 hPa	31 km globally, 62km for the Ensemble of Data Assimilations (EDA), 137 levels to 0.01 hPa
Output frequency (temporal resolution)	6-hourly analysis fields Forecast fields on surface and pressure levels 3-hourly up to 24 hours, with reduced frequency up to 10 days	Hourly analysis fields, 3-hourly for the Ensemble of Data Assimilations (EDA) Hourly forecast fields, 3-hourly for the Ensemble of Data Assimilations (EDA), up to 18 hours, with reduced frequency up to 10 days (not in initial release)
Uncertainty estimates	None	From a 10-member Ensemble of Data Assimilations (EDA) at 63 km resolution
Model input	As in operations (inconsistent SST)	Appropriate for climate (e.g. CMIP5 greenhouse gases, volcanic eruptions, SST and sea-ice cover)
Input observations	As in ERA-40 and from Global Telecommunication System	In addition, various newly reprocessed datasets and recent instruments that could not be ingested in ERA-Interim
Variational bias scheme	Satellite radiances	Also, ozone, aircraft and surface pressure data
Satellite data	RTTOV-7, clear-sky, 1D-VAR rainy radiances	RTTOV-11, all-sky for various components
New parameters	ERA-Interim contains about 100 parameters on surface and single level alone, plus parameters on other level types.	ERA5 contains over 240 parameters on surface and single level alone, plus parameters on other level types. For specific parameters please compare the technical documentation linked below.
Parameters removed		A few parameters present in ERA-Interim are not available in ERA5. For availability of specific parameters please see the technical documentation linked below.
Additional innovations		Long-term evolution of CO ₂ in RTTOV, cell-pressure correction SSU, improved bias correction for RSs, EDA perturbations for sea-ice cover

Table 2.8: *Main differences between ERA5 and ERA-I (ECMWF 2020d).*

Although, many studies find a significant improvement in the performance of ERA5 over ERA-I in resolving the atmospheric stratification (GRAHAM et al. 2019; TETZNER et al. 2019; PALARZ et al. 2020) and mention the immanent replacement of ERA-I by ERA5, it is still of great importance to evaluate its quality in more detail. ERA-I has been widely used

over recent years to study the stratification of the atmosphere (PARLARZ et al. 2018; SHAHI et al. 2020; ZHANG et al. 2011) and has shaped our present knowledge about temperature inversions. The evaluation of ERA-I’s limitations and capabilities using high quality in-situ measurements will improve our ability to interpret results from earlier studies.

An important point to be noted is that in ERA-I and ERA5 the height of the land surface does not represent the geometric altitude, but rather the geopotential height (ECMWF 2020c). The surface geopotential elevation of the grid point used is obtained following the method outlined in ECMWF (2020c): the surface geopotential is divided by the acceleration due to gravity ($g = 9.80665\text{m/s}^2$) to give the geopotential height in metres a.s.l. The extracted value for surface height is 428m a.s.l. for the ERA-I model and 151m a.s.l. for ERA5.

The limitations of ERA-I and ERA5 data in reproducing temperature inversions are examined in two ways. The temperature deviations between UAV and reanalysis data at the same elevations are calculated to discover elevations with high and low agreement. Finally, the ability of the reanalysis data to reproduce atmospheric stratification is evaluated. This is achieved by comparing the results of the calculated inversion statistics with the UAV data.

2.4 Additional data

One aim of this study was to look at the interrelation between the atmospheric stratification and the ablation of the Mittivakkat Glacier. In order to quantify this, additional data was collected. For example, data on water temperature and water level were collected in a proglacial lake. The lake was located between locations D and E and was fed by the ablation of the Mittivakkat glacier (Figure 2.11). The measurements were performed using LevelStick, a pressure sensor with an automatic logging function manufactured by the Dutch company Leiderdorp (Table 2.9) (LEIDERDORP 2020).

Location	Resolution	Accuracy	Time interval
65°41' N, 37°52' W	0.1hPa	$\pm 0.1^\circ\text{C}$	1 min

Table 2.9: *Pressure sensor and measurement specifications (LEIDERDORP 2020).*

This sensor calculates the water level from the hydrostatic pressure of the water column above the sensor. The logged data can be retrieved from the sensor with a freely available smartphone app called LevelStick. This app can be used to set the log interval, to see real-time measured values, to define the settings, to send the data, or to retrieve and view the logged data. The communication between sensor and smartphone was by Bluetooth. The sensor cable was fixed around a boulder and the sensor installed approximately one metre below the surface of the lake. The measurements were begun at 7pm [UTC-2] on 8 July 2019 and ended on 18 July 2019. The post-processing of the data involves averaging over

a time interval of $t = 1$ hour and the conversion from hPa to cm, where a 100hPa change corresponds to a 102cm water level change.

Figure 2.11: On the left is the automatic data logger pressure sensor LevelStick. On the right is the proglacial lake located between locations D and E (red dots). The blue dot shows the location where the sensor was placed (LEIDERDORP 2019).

As a further data source, the water level of a proglacial river in the Mittivakkat valley was measured (Figure 2.12) $65^{\circ}40' N$, $37^{\circ}54' W$), a location already used in other studies (HASHOLT & MERNILD 2006). The site was marked by an angular boulder whose top rose above the surface of the water. The distance from the upper edge of the boulder to the water surface was measured every day when we passed on our way up and down the transect. The measurements therefore were not carried out at precisely the same time every day.

Figure 2.12: The level of the proglacial river was quantified using a measuring stick from the edge of the boulder (reference point) to the water surface (left side). The site (blue dot) was close to location B (red dot).

During the field study, the ablation of snow and glacier ice was measured using a network of ablation stakes on the Mittivakkat glacier. The network was installed by the University of Copenhagen for mass balance measurements of the Mittivakkat glacier. The network consists of several stakes (Figure 2.13) which are drilled 5m into the surface of the glacier. The distance between the individual stakes is approximately 200m-250m (KNUDSEN & HASHOLT 2004; HASHOLT et al. 2016). Measurements were made at the individual stakes along the transect. The ablation rate is defined as the change in the length of stake exposed above the glacier surface. The locations of the stakes used in this study are shown in Figure 2.1 and the names of the stakes visited and dates of the visits in Figure 2.13.

Figure 2.13: The left side lists the names of the ablation stakes and dates when measurements were taken. A typical ablation stake is shown in the centre and the quantification method on the right.

2.5 Data processing

Processing the UAV data

The iMET data collected during the field campaign was saved in text files. Each file contains the atmospheric profiles from a different day. In order to isolate the data from the ascents, the data from the descents and data collected before take-off were identified and removed. The beg starting point of the descent can be derived from the maximum flight height, and the pre-take-off data obtained from the notes in the field book. The data was then averaged over height intervals of $z = 12\text{m} \pm 1\text{m}$, which corresponds to the vertical accuracy of the on-board GPS of the iMET-XQ2 sensor. The comparison between the original (blue) and the averaged data (red) is shown in Figure 2.14. It can be seen that the averaged data still follow closely the course of the original data sets.

Figure 2.14: Comparison of six original profiles (blue) with averaged UAV profiles (red).

Atmospheric lapse rate calculation

The lapse rate (LR), which is the rate at which temperature changes with altitude in the atmosphere, was calculated for every profile. The change of air temperature with height at any location and time is called the normal or environmental lapse rate (ELR) (Figure 2.15). How much the temperature varies in the vertical depends on several factors (e.g., condensation, radiation, convection) and changes from day to day and hour to hour, but is on average 0.65°C per 100m (ACKERMAN & KNOX 2012; BRITANNICA 2016).

The adiabatic lapse rate (processes in which no heat or mass is lost or gained by the system) describes how the temperature of a parcel of air changes due to sinking and rising processes. There are two different adiabatic lapse rates: dry and moist (Figure 2.15) (BRITANNICA 2016; WAUGH 2002).

When a parcel of unsaturated air (relative humidity $< 100\%$) rises adiabatically, the air expands because pressure decreases with height. Consequently, the temperature of the air parcel will fall. If the air is dry, the rate of temperature change will be 1°C per 100m. This is called the dry adiabatic LR. When the parcel of air cools to its dew point temperature (relative humidity = 100%), condensation occurs, which releases latent heat. The release of the latent heat will in turn slow down the cooling. The moist adiabatic LR depends on the amount of water and vapour present in the air and is therefore highly variable. It is taken to average 0.6°C per 100m (BRITANNICA 2016; WAUGH 2002).

The difference between the ELR and either the moist or dry adiabatic lapse rate determines the stability of the atmosphere (Figure 2.15). An atmosphere is unstable when during uplift an air parcel cools less quickly than the ambient air. The air parcel thus remains warmer relative to the ambient air and the associated lower density leads to further uplift of the air parcel. If the ambient air cools faster than a dry-adiabatic lifted air parcel (1°C per 100m), the stratification is called dry-labile. If the ambient air cools

Figure 2.15: Temperature profile in a stable (a) and unstable (b) atmosphere.

faster than a moist-adiabatic lifted air parcel (0.6°C per 100m), the stratification is called moist-labile. Stable conditions occur when the lifted air package cools faster than the ambient air. The air package then has a higher density than the ambient air and is blocked from

rising further. When the air is lifted dry-adiabatically and cools faster than the ambient air, this is known as dry-stable stratification. When a moist-adiabatically lifted air package cools faster than the ambient air, it is called moisture-stable stratification. The atmosphere is neutral if both LR are equal (BRITANNICA 2016; HUPFER & KUTTLER 2006). The LR was calculated (height intervals $z = 100\text{m}$) using the formula given below (BRASSEUR & JACOB 2017):

$$(2.1) \quad \Gamma = - \left(\frac{dT}{dz} \right)$$

where Γ is the LR, dT is the temperature difference and dz the height difference. LR values were calculated for 100m intervals: the value at 100m a.g.l. was subtracted from the value at 0m a.g.l.; the value at 200m a.g.l. from that at 100m a.g.l., and so on, to determine the type of stratification present in each 100m layer. It should be noted that the LR was only calculated for whole 100m steps. For example, if the profile only went up to 450m a.g.l., the LR would be calculated up to only 400m. This is also the reason why the profiles shown often go up to only 400m a.g.l.

Figure 2.16: The original UAV profiles are indicated in black, using only the temperature values at 0m, 100m, 200m, 300m, 400m and (where available) 500m. Alongside are the same profiles with the individual 100m segments colour-coded for the calculated LR of the respective segment.

The original UAV profiles were then plotted, showing the temperature values only at 0m, 100m, 200m, 300m, 400m and 500m a.g.l. to display the general temperature trend at a particular time of day. The individual 100m segments are colour-coded according to the LR value calculated for each layer. Thus the 100m segment is coloured red if the layer is stable; green, if the layer is neutral; blue, if the layer is unstable (Figure 2.16). The exact point at which stratification becomes stable, neutral, or unstable has been taken from the literature (HUPFER & KUTTLER 2006: dry adiabatic LR).

$$\Gamma \leq -1 = \text{Unstable} \quad \Gamma = 0 = \text{Neutral} \quad \Gamma \geq 1 = \text{Stable}$$

A number of different definitions of positive or negative lapse rate appear in the literature. In this study, lapse rates are taken as positive when temperature decreases as elevation increases (GARDNER et al. 2009).

Spatial-temporal interpolation

The profiles collected in the course of the transect were divided into two groups: those collected during the walk up and those from the walk down. These two sets of profiles had to be interpolated on a daily basis in order to make the atmospheric pattern along the described transect and during the given time frame more visible. To achieve such an interpolation, the elevations (and related atmospheric variables) of all profiles were recalculated from m a.s.l. to m a.g.l., then a grid was created with defined x and y values, where x is the distance from location B (0m-6166m) (Figure 2.17 a) and y is the height above ground level (0m-500m a.g.l.).

Figure 2.17: *The individual steps of the interpolation. a) how the distance values were obtained, b) the individual profiles of a transect in the previously created grid, c) the interpolated grid, d) how the profile was lifted back to the original heights and e) the finished profile. The red dashed line between 1742 and 2119 shows the approximate starting point of the glacier.*

The grid consists of 500 columns and 500 rows. The atmospheric variables measured by the UAV (air temperature, humidity, and pressure) are transferred into the grid (z coordinate) (Figure 2.17 b). This grid is the environment in which the interpolation is carried out. The empty cells are filled with values calculated using linear interpolation (Figure 2.17 c). Using this method, neighbouring data points are connected with a straight line and every intermediate value (cell) is derived from the line. When the interpolation is complete, the interpolated profile is lifted to match the original topography. The elevation at each location corresponds to the elevation information in m a.s.l. from the original profile acquired at the corresponding location (Figure 2.17 d). Profiles B to H were used in the interpolations, A was excluded to show a relatively straight transect from the sea to the nunatak.

In addition to the temperature profiles, the potential temperature of each profile is calculated and interpolated as a transect. The potential temperature provides information about the degree of stability or lability of the air masses. It is defined as the temperature that a parcel of air at height z would have if it were brought adiabatically (without energy exchange with the environment) to a pressure level of 1000hPa. This pressure level is approximately equal to that at sea level and therefore represents a suitable reference value. The potential temperature would increase with altitude in a stable stratification; remain constant in a neutral stratification; and decrease in an unstable stratification (CHROMOW 1940). The potential temperature Θ was calculated for every profile using the formula (CHROMOW 1940)

$$(2.2) \quad \Theta = T \left(\frac{P_0}{P} \right)^{0.2884}$$

where T is the temperature measured in Kelvin, P is the pressure at the location where the measurement was taken, P_0 is the standard reference pressure (1000 hPa), and 0.2884 is the gas constant of air divided by the specific heat capacity of air at constant pressure.

Figure 2.18: Temperature variation at the different AWS during the time period when the profiles were collected. 25m a.s.l. represents COAST, 196m a.s.l. 196, 440m a.s.l. GEUS and 748m a.s.l. CCMORE.

The diurnal variations in temperature in this region and at this time of the year are rather small and were even smaller during the period in which we collected the profiles along the individual transect (Figure 2.18). For this reason, we concluded that, even though the interpolated profiles were not collected at the same time of day (apart from B & C and D & E), we could interpolate and show them on one plot.

Data point reduction

A comparison of UAV, RS, and reanalysis data requires, as a first step, taking temperature measurements at the same height levels. Figure 2.19 shows UAV, RS, ERA5 and ERA-I profiles, all representing the condition of the atmosphere at the same time on the same day. The red dots represent the measurement points, i.e., the heights at which, for example, temperature values were measured. This clearly shows how much the measurement heights differ.

Figure 2.19: UAV, RS, ERA5 and ERA-I profiles, all representing the situation of the atmosphere at the same time. The red dots are the measurement points.

To find common measuring points, the original unaveraged UAV profiles were compared to contemporaneous RS, ERA5, and ERA-I profiles. The advantage of using the original UAV profiles is their measurement frequency (every second). This allows for the use of the same height or pressure levels. The result was a new data set with common heights including the corresponding temperature values from the UAV, RS, and reanalysis data. Finally, the RS and reanalysis data were subtracted from the UAV to calculate the deviation.

2.6 Temperature inversion detection algorithm

An algorithm to detect temperature inversion was developed to identify temperature inversion characteristics objectively over all the datasets (UAV, RS, ERA5 and ERA-I). All profiles were scanned from ground level (0m a.g.l.) to the top (500m a.g.l.) to identify whether or not there is an inversion, and, if so, which inversion type (SBI or EI). The inversion base (Z_{Base}), inversion top (Z_{Top}), inversion thickness (ΔZ), inversion strength (ΔT), and inversion frequency (f , percentage of days on which temperature inversions were recorded) (Figure 2.20) are also included.

Figure 2.20: *Detected inversion characteristics.*

To ensure that the temperature inversion detection algorithm is not detecting artificial inversions due to measurement (sensor) inaccuracy, temperature increases between adjacent levels of less than 0.3°C in the UAV data were ignored. This takes into account the absolute accuracy of the temperature sensor employed (INTERMET 2018). This high degree of accuracy was confirmed by a test carried out in the field (2.3.1.). According to instrument manufacturer Vaisala (2020), the typical temperature uncertainty for the Vaisala sonde type RS41, which was used during the field study at the station in Tasiilaq, is 0.3°C . Thus, the onset of an inversion in the RS data can be detected if there is a temperature increase of at least 0.3°C between two levels.

In other studies dealing with temperature inversions and using ERA5 or ERA-I data, there had to be a temperature increase greater than 0°C between two levels for the layer to be identified as an inversion (SHAHI et al. 2020; PALARZ et al. 2018, 2020; ZHANG et al. 2011). This means that the onset of a temperature inversion has been detected if the temperature increase between two levels is larger than 0°C . In order to compare our reanalysis inversion statistics results with these studies, this threshold ($>0^{\circ}\text{C}$) was also applied to detect inversions in the reanalysis data. In addition, to see how the inversion statistics change, a second run with a threshold of 0.3°C was performed.

Here, Z_{Base} is the elevation level at which the temperature starts to increase with height. Z_{Top} is the top of the inversion, which is the elevation level at which the temperature starts to decrease. ΔZ is the difference in altitude between Z_{Base} and Z_{Top} . The difference between the temperature at Z_{Base} and Z_{Top} determines ΔT . Thin layers ($\leq 100\text{m}$) of negative lapse rates between two detected inversion layers are ignored and considered to be part of the same inversion layer. Other studies include non-inversion layers of 100m thickness (KAHL 1990; BRADLEY et al. 1992; GILSON 2018; SHAHI et al. 2020; ZHANG et al. 2011).

Figure 2.21: The blue profile with the red dots is the UAV temperature profile with the measurement points. The area highlighted in light red is the inversion thickness found by the algorithm. This shows clearly how the different threshold values (the tolerated thickness of the layer of negative lapse rate between two inversion layers: 0, 20, 50 and 100m) affect the inversion characteristics.

layer atmosphere. For this reason, and to follow precedent and recommendations in the literature for a 100m or lower thickness threshold for non-inversion layers (KAHL 1990; GILSON 2008; SHAHI et al. 2020; ZHANG et al. 2011), our study accepted a 100m or less non-inversion layer embedded in two inversion layers (ZHANG et al. 2011).

After identifying all the inversion characteristics mentioned, a new dataset was created to include all the quantified characteristics. In addition, a visual representation of the characteristics for every profile was created in order to have a visual check on how well the algorithm works and to detect possible errors. The visual output of the algorithm is shown in Figure 2.19.

The sensitivity of the temperature inversion statistics to changes in the tolerated thickness of the layer of negative lapse rate between two inversion layers was tested. The tested thicknesses were 0m, 20m, 50m and 100m. Approximately 28% and 25% more inversions were detected with thickness thresholds of 0m and 20m respectively compared to a thickness threshold of 100m. Due to the high resolution of our input data (12m) the thresholds of 0m and 20m for non-inversion layers give a fragmented picture of the inversion (Figure 2.21). Both of the higher threshold values for negative lapse rate layer thickness (50m and 100m) can capture the inversion and show a clearer picture of the boundary

RESULTS AND DISCUSSION

The results of this study will be presented and discussed in this chapter. Results and their discussion are usually treated in separate chapters. However, as this study deals with so many different, yet related, aspects of temperature inversions in Ammassalik, to separate results and discussion would make it harder for the reader to remember the context when previous findings are evaluated and discussed.

This chapter is divided into four sections. In 3.1 the high-resolution UAV profiles are compared with RS, ERA5, and ERA-I profiles. In 3.2 the inversion characteristics above the different surface types and along the transect are presented. In 3.3 the effects on the activity and intensity of the temperature inversions of synoptic scale events observed during the field campaign are shown. Finally, in 3.4, the impact of temperature inversion statistics on the ablation of the Mittivakkat glacier is presented. Unless otherwise stated, all dates and times are in UTC-2.

3.1 Temperature characteristics of different data sets at near-surface levels

The limitations and advantages of the RS, ERA5, and ERA-I datasets, which are commonly used in temperature inversion studies, are discussed in this chapter. The ability of these datasets to represent temperature inversions realistically is determined on the basis of two criteria. Firstly, how well do the temperature profiles of the RS, ERA5, and ERA-I data match those of the UAV data (3.1.1). Secondly, how different are the inversion statistics of the RS, ERA5, and ERA-I data from the UAV data (3.1.2). To obtain this information, 21 UAV profiles collected at location A are compared with 21 simultaneously recorded RS profiles from the Tasiilaq station and 21 ERA5 and ERA-I profiles.

3.1.1 Vertical temperature profiles

3.1.1.1 Radiosonde

Figure 3.1 illustrates the temperature profiles from all 21 RS and UAV datasets. Overall, the greatest discrepancy between the two datasets can be observed in the first 100m a.g.l. The profiles become more similar with increasing height, but the degree of agreement varies

Figure 3.1: UAV profiles (black) and RS profiles (red), per day and observation (10:00 and 22:00). The black data show the profiles at 10:00 (AM) and the red ones at 22:00 (PM). The temperature at the nearby AWS SYNOP is shown in green.

across the observations. The correlation between RS and UAV temperature values measured at the same heights is calculated to evaluate the degree to which they agree with each other.

Data	RMSE [°C]	r
All	3.28	0.49
AM	3.55	0.34
PM	2.93	0.62

Table 3.1: RMSE and the correlation coefficient between UAV and RS data.

If both profiles moved strictly in the same direction, a Pearson correlation coefficient (r) of 1 and a root-mean square error (RMSE) of 0°C would be obtained. However, if we compare the temperature values from all RS and UAV profiles, we observe a high RMSE of 3.28°C and a weak correlation of $r = 0.49$. The RMSE is slightly lower, and the r value slightly higher in the PM profiles if we separate the data into 10:00 (AM) and 22:00 (PM) profiles (Table 3.1).

z [m a.g.l.]	RMSE [°C]	r
0-100	3.55	0.33
100-200	1.45	0.93
200-300	1.83	0.93
300-400	2.17	0.79
400-500	1.89	0.69

Table 3.2: RMSE and correlation coefficient per 100m elevation band.

Figure 3.2: The correlation between the RS and UAV profiles at AM (left side) and the correlation between the two data sets at PM (right side). Different 100m intervals are plotted with different shapes and colors.

We must also always consider that the two datasets were obtained approximately 15km apart. There are various local factors (e.g., surface topography and local wind systems) that affect the air temperature lapse rates in Tasiilaq and Sermilik. With increasing height these local factors should have less impact on the measured values. Figure 3.2 and Table 3.2 show the degree of agreement between both data types in different 100m elevation bands. The elevation band 400m to 500m a.g.l. is excluded due to the low number of common data points (3 common data points). The number of common data points varies from band to band and depends on the number of mandatory pressure levels in the respective elevation bands.

As expected, the correlation is strong and positive and the RMSE moderate to low in all elevation bands higher than 100m a.g.l. There is almost no correlation and a high RMSE below 100m a.g.l. Figure 3.3 shows that the greatest difference between both data types within the 100m a.g.l. elevation band can be observed at the surface. The RS data show lower temperature values than the UAV data. This temperature difference at the surface could be due to a RS cold bias.

Figure 3.3: Temperature values of the RS profiles were subtracted from those of the UAV profiles at the same altitude (UAV minus RS). The boxplots show the deviation across the bottom in 10m steps. The thick black line in the boxplot is the median. The black dot is the average deviation per altitude. The red dots above the boxplots show the distribution of the deviations in °C and the number of common data points per altitude class is shown to the right of the boxplot.

Synoptic observational data of temperature recorded at SYNOP station were used to check if there is a cold bias in the surface observations of the RS. In order to do so, temperature values of the RS surface level at 0m a.g.l. were compared with the temperature data from SYNOP collected at the same height, time (10:00 and 22:00), and location. This means, as a corollary, that both should show similar values. This quality-assessment approach to the surface RS observations has already been used in other studies such as WANG & YOUNG (2005). The comparison revealed a positive correlation close to 1 ($r = 0.89$), but also showed a high RMSE value of 5.02°C (Figure 3.4). This means that both datasets follow a similar trend, but most of the time the RS temperature values are, except for one profile, lower than those from the AWS.

Figure 3.4: *RMSE and correlation between temperature values of the RS at surface level (52m a.s.l.) and of SYNOP.*

We assume that the data from the SYNOP station are more reliable than the surface level observations of the RS. In order to see how the correlation would change using the more reliable data, the surface level values from the surface level observation from the RS at 0m a.g.l. was replaced by the corresponding SYNOP station values and the correlation calculation revised. The results can be seen in Figure 3.5 and Table 3.3. The RSME was reduced to 1.91°C in the PM data and to 2.23°C in the AM data. The correlation coefficient increased to $r = 0.84$ in the PM data and to $r = 0.72$ in the AM data.

Data	RMSE [$^\circ\text{C}$]	r
All	2.08 (- 1.20)	0.78 (+ 0.29)
AM	2.23 (- 1.32)	0.72 (+ 0.38)
PM	1.91 (1.02)	0.84 (+ 0.22)

Table 3.3: *RMSE and the correlation coefficient between UAV and corrected RS data.*

Figure 3.5: *Correlation between the UAV and RS data are given on the left and for the AM and PM data on the right. The points of the 100m intervals have different shapes and colors.*

3.1.1.2 ERA5 and ERA-I Model Level

The data presented in Figure 3.6 show that the profiles from both reanalysis products differ most from the UAV data at levels close to the surface. These differences occur not only between the UAV and the reanalysis products, but also between the reanalysis products themselves. For most of the time, ERA-I seems to record colder temperatures at the surface than ERA5. There is more agreement between all three datasets above the near-surface, but the degree of agreement differs from day to day.

Figure 3.6: UAV profiles (black), RS profiles (red), ERA5 profiles (green) and ERA-I profiles (blue), by day and observation time (10:00 and 22:00). The black data show the profiles at 10:00 (AM) and the red ones at 22:00 (PM).

In addition, both reanalysis products differ from the RS profiles, although the RS profiles are taken into account in both reanalysis products. The correlation between both reanalysis products and UAV profiles have been calculated. Figure 3.7 and Table 3.4 show the correlations of all UAV and reanalysis products grouped into AM (10:00) and PM (22:00) datasets. ERA-I shows a high RMSE of 3.39°C in the AM data and a weak correlation of 0.28 (Table 3.4). The RMSE of 3.17°C is only slightly better in the PM data, but the relationship between both datasets improved by 0.30 (0.58). Overall, ERA5 performs better, especially in the PM dataset with an RSME of 1.82°C and a high correlation coefficient of 0.88, suggesting that there is an improvement in the representation of the lower atmospheric temperature in the ERA5 data. The PM data performed better in both reanalysis datasets, which fits well with the RS observations.

As in the case of the RS data, there is a greater difference between UAV and both reanalysis products in the first 100m a.g.l. and a decreasing difference with increasing height (Figure 3.7). This is reflected in the RSME and the values of the correlation coefficients (Table 3.5). The RSME is highest and r values lowest in the first elevation band in the ERA-I and ERA5 data, but the RSME becomes lower and the r value higher in the second, third, and fourth elevation bands. Because of the lower vertical resolution, no common data points could be found in the ERA-I and UAV data between 400 and 500m a.g.l.

Figure 3.7: At the top is the correlation between ERA-I and UAV profiles AM (top left) and PM (top right). Below is the correlation between ERA5 and the UAV profiles AM (bottom left) and PM (bottom right).

Data	RMSE [°C]	r
ERA-I		
All	3.29	0.45
AM	3.39	0.28
PM	3.17	0.58
ERA5		
All	2.12	0.81
AM	2.36	0.72
PM	1.82	0.88

Table 3.4: RMSE and the correlation coefficient between the UAV and reanalysis products.

ERA-I			ERA5		
z [m a.s.l.]	RMSE [°C]	r	z [m a.s.l.]	RMSE [°C]	r
0-100	4.12	0.18	0-100	2.54	0.65
100-200	1.92	0.82	100-200	2.16	0.88
200-300	1.73	0.87	200-300	1.77	0.90
300-400	1.37	0.95	300-400	1.42	0.92
400-500			400-500	0.76	0.97

Table 3.5: RMSE and the correlation coefficient between the UAV and reanalysis products per 100m altitude class.

Although ERA5 is marginally less accurate than ERA-I in the elevation bands above 100m a.g.l., ERA5 clearly performs better than ERA-I in the first 100m. In addition, as Figure 3.8 shows, there is a cold bias in the ERA-I data, especially at the height level 0-100m a.g.l. (Figure 3.8 1a). The ERA5 temperatures between 100 and 300m a.g.l tend to be warmer than in the UAV data followed by colder temperature data in the ERA-I data between 300m a.g.l. and 400m a.g.l. There is also a warm bias in the first 100m a.g.l. of the ERA-5 data, which

decreases with elevation (Figure 3.8 2a). The differences in the first 100m a.g.l. between the two reanalysis products are shown in Figure 3.8 (1b & 2b). The detected cold bias in ERA-I is highest at the surface (0m a.g.l.) (Figure 3.8 2b). The differences and the cold bias are already smaller at 2m a.g.l., which indicates a temperature increase from 0 to 2m a.g.l.

Figure 3.8: Temperature values of the (1a) ERA-I and (2a) ERA5 data were subtracted from those of the UAV profiles at the same altitude. The boxplots show the deviation per 100m altitude class. The thick black line in the boxplot is the median. The black dot is the average deviation per altitude. The red dots above the boxplots show the distribution of deviations in °C. The number of common data points per altitude class is shown to the right of the boxplot. On the right, the temperature values of (1b) ERA-I and (2b) ERA5 were subtracted from those of the UAV profiles at the same altitude. The altitudes were given by the reanalysis products and the altitude of their model levels.

The warm bias in ERA5 remains relatively constant in the first 100m a.g.l. (Figure 3.8 2b). It is only the temperature values at 2m a.g.l. that are slightly colder. The consequence is that the temperature increases in the levels above 2m a.g.l. These differences could reflect the more complex atmospheric processes at the near-surface which cannot entirely be simulated by the models.

There are other studies which have observed cold and warm biases close to the surface in ERA-I and ERA5 data. TETZNER et al. (2019) evaluated the accuracy of different meteorological parameters in ERA-I and ERA5 by comparing them with in-situ observations from dif-

ferent AWS in Antarctica at 2m a.g.l. They found a warm bias in the interior of Antarctica and a cold bias along the surface coast of the southern Antarctic peninsula in both reanalysis products at this elevation, with a slight reduction in the ERA5 data. JONES & LISTER (2015) also compared near-surface air temperature values in Antarctica recorded between 1979 and 2013 from 40 AWS with ERA-I and found a warm bias inland and a cold bias at some of the coastal locations. JONES et al. (2016) compared different meteorological

variables in the ERA-I with different AWS, RS data, and meteorological data collected on a research vessel at points in the Amundsen sea embayment in Antarctica. The authors found a cold bias along the coast in the 2m a.g.l. temperature field in ERA-I. GRAHAM et al. (2019) used meteorological variables from RS launches in the Frame Strait to evaluate the performance of ERA-I and ERA5 data. They observed a cold bias for ERA-I and a warm bias for ERA5 at the surface during summer.

One major cause of such biases could be that already mentioned in 2.3.3, i.e., the difference in the surface elevation between the ERA5 and ERA-I data and the observational data (UAV and AWS). Surface heights of 428m for ERA-I m a.s.l and 151m a.s.l for ERA5 were calculated (2.3.3.). These differences in elevation could explain the cold bias in ERA-I. Such a cold bias due to elevation dissimilarities between ERA5 and ERA-I data and observational data was also reported by other studies (LUO et al. 2019; DUTRA et al. 2020).

3.1.2 Inversion characteristics

3.1.2.1 Radiosonde

Temperature inversions were detected in 17 out of 21 UAV observations and in 20 of the RS observations (Table 3.6). 14 more EI were detected in the UAV data compared to the RS data. Exactly the opposite holds in the case of SBI. SBI were observed in 20 RS but none were recorded in the UAV profiles.

Data	Relative fraction	No. of profiles with inversions	No. of SBI	No. of EI
UAV	81 %	17	0	20
RS	95 %	20	20	6
Corrected				
RS	66%	14	5	11

Table 3.6: Inversion statistics for UAV and RS data.

One reason for the high numbers of SBI, and for the temperature inversions in general found in the RS observations, could be the cold bias that was detected in the RS surface observations (Figure 3.9). The number of SBI is reduced to 5 observations and the number of EI increases to 11 observations if we replace the surface temperature value of the RS profiles with the corresponding AWS value.

Figure 3.9: The first 50m of a few a) RS and b) RS profiles where the first measurement at 0m a.g.l. was replaced by the temperature value from AWS SYNOP. Blue lines represent the profiles, and red dots the measurement points. The area highlighted in light red is the inversion found by the algorithm.

Figure 3.10 illustrates the variability and difference in the detected inversion characteristics (Z_{Base} , Z_{Top} , ΔZ and ΔT) in the AM (Figure 3.10, black dates) and PM (Figure 3.10, red dates) UAV, RS, and corrected RS (named RS2 - first observation replaced with the corresponding

AWS value) profiles. The constant low temperature at the surface level in the RS profiles triggers high ΔT and ΔZ values (for example, on 13.07.2019, 14.07.2019, and 18.07.2019 PM). Although still very different, the characteristics in the RS2 profiles are somewhat more similar to those of the UAV profiles than those of the RS profiles. The coarse vertical resolution of the RS data precludes a realistic representation of thin inversion layers, thus many EI were sampled as part of an SBI and separate EI layers were often counted as one. The observed limitations of RS soundings for temperature inversion studies are in good agreement with previous studies (PALO et al. 2017; SKONY et al. 1994; VIHMA et al. 2011).

Figure 3.10: The inversion characteristics of the UAV and RS data and the RS data where the first value was replaced with that of the AWS SYNOP (RS2). The bars symbolize the inversion, the bottom of the bar is the inversion base in m a.g.l., the top of the bar is the inversion top in m a.g.l. The strength is shown by colour: the higher the temperature difference inside the inversion, the warmer the colour. The black data represents the profiles at 10:00 (AM) and the red the profiles at 22:00 (PM).

3.1.2.2 ERA5 and ERA-I Model Level

Temperature inversions are found in all 21 ERA-I and ERA5 profiles (Table 3.7). One reason for the high number of inversions found could be the lower temperature increase required between two measurement points (more than 0°C). For the RS and UAV data, a temperature increase of more than 0.3°C between two measurement points had to be present. 19 instead of 21 inversions are detected if we change this threshold (required temperature increase between two measuring points) from greater than 0°C to greater than 0.3°C in the ERA-5 data while the number of inversions in the ERA-I data remains the same.

Data	Relative fraction	No. of profiles with inversions	No. of SBI	No. of EI
Threshold of $> 0^{\circ}\text{C}$				
ERA-I	100 %	21	21	9
ERA5	100 %	21	8	21
Threshold of $\geq 0.3^{\circ}\text{C}$				
ERA-I	100 %	21	21	5
ERA5	90 %	19	6	18

Table 3.7: Inversion statistics for ERA-I and ERA5 data.

Figure 3.11: The first 50m a.g.l. of a sample of profiles: ERA-I (top) and ERA5 (bottom) to show the cold bias in ERA-I with the resulting high number of SBI on the one hand and the cold bias of the ERA5 in 2m a.g.l. with the resulting high number of EI on the other. Blue lines represent the profiles, and the red dots the measurement points.

A higher number of SBI and a lower number of EI were found in the ERA-I profiles compared to the ERA5 and UAV profiles. Even though we can observe a reduction in detected SBI in the ERA5 data, there are still more than in the UAV data. This does not change even if the threshold is raised to $\geq 0.3^{\circ}\text{C}$ in the inversion detection algorithm (Table 3.7). The number

of EI in the ERA5 data is higher than in the UAV data, but we obtain a lower number if we apply the higher threshold of $\geq 0.3^\circ\text{C}$. In our study, the cold bias detected at the surface of the ERA-I profiles explains the high number of SBI in this dataset (Figure 3.11). The warm bias in ERA5 remains relatively constant in the first 100m a.g.l., with only the temperature values at 2m a.g.l. slightly colder, which explains the small number of SBI (temperature decrease between 0 - 2m a.g.l.) and high number of EI (temperature increase from 2m to 10m a.g.l.) detected in the ERA5 profiles (Figure 3.11).

Figure 3.12: Inversion characteristics of UAV, ERA5 and ERA-I data. The bars symbolize the inversion. The bottom of the bar indicates the inversion base in m a.g.l., the top of the bar the inversion top in m a.g.l. The strength is shown by colour: the higher the temperature difference, the warmer the colour. For the ERA5 and ERA-I data framed in red, a temperature increase of more than or 0.3°C between two measurement points was required for the detection of an inversion, for the unframed data a temperature increase of more than 0°C . The black data show the profiles at 10:00 (AM) and the red at 22:00 (PM).

This high number of summer SBI observed in ERA-I and ERA5 and low number of summer EI in ERA-I is in keeping with other studies. PALARZ et al. (2020) analyzed the uncertainties in the characteristics of temperature inversion in the lower troposphere among different reanalysis products (including ERA-I and ERA5) by comparing them with upper-air soundings above Europe. They found that the vast majority of reanalysis products overestimated SBI and underestimated EI. ZHANG et al. (2011) followed a similar approach. They compared ERA-I data with RS observations in the Arctic and Antarctic and found an overestimation of SBI in ERA-I data at most stations.

The differences between the inversion characteristics (Z_{Base} , Z_{Top} , ΔZ and ΔT) in the AM (Figure 3.12, black dates) and PM (Figure 3.12, red dates) of the UAV, ERA5, and ERA-I profiles are shown in Figure 3.12. The degree of correspondence between the UAV detected and ERA-I inversion characteristics is low. In most cases the SBI ΔT detected in the ERA-I data is very high, with values up to 12°C (e.g., Figure 3.12, 16.07.2019 PM). This overestimation of ΔT and ΔZ in ERA-I was also observed in the study conducted by ZHANG et al. (2011), where they compared observational with ERA-I and modelled data. In our study, the high ΔT values are due to the (already mentioned) low temperature values near the surface in the ERA-I datasets. Such low surface temperatures trigger the detection of SBI, whereas failure to note layers of up to 100m thick with negative lapse rate coupled with the low vertical resolution, could result in some of the EI being sampled as part of an SBI and thus enhancing inversion thickness. Individual EI layers were often counted as one due to reduced vertical resolution with increasing height (10 levels). Both limitations resulted in an overestimation of ΔZ in the ERA-I dataset (e.g., Figure 3.12 08.07.2019, 14.07.2019, 17.07.2019, 18.07.2019 AM). The agreement between the inversion characteristics of UAV, ERA5, and ERA-I does not increase significantly when the required temperature increase between two measurement points is changed from greater than 0°C to greater than or equal to 0.3°C (Figure 3.12, red squares).

The inclusion of a larger number of vertical levels (a total of 16 levels) and the improvements implemented in physical parametrization schemes in ERA5 have helped to reduce the overestimation of ΔZ (PALARZ et al. 2020). Overall, the performances of ERA5 and ERA-I reveal that there is only a small improvement in representation of temperature inversions in ERA5 data.

3.2 Impact of surface types and settings on temperature inversions

Subsection 3.2.1 of this chapter describes the spatial and temporal variations of temperature inversion characteristics (Z_{Base} , Z_{Top} , ΔZ , ΔT and f) above 8 different locations and four different surface types on Ammassalik Island. Subsection 3.2.2 focuses on the horizontal temperature pattern above the different surface types of the transect.

3.2.1 Vertical profiles

The analysis of 113 UAV profiles reveals that temperature inversions are frequent in the study area in summer. Overall, temperature inversions were found in 83% of all UAV profiles (Table 3.8, Figure 3.13). This compares well with values reported in other studies conducted in east Greenland during summer. GILSON et al. (2018) analysed more than 22,000 soundings from three east Greenland coastal stations during the melt season and found temperature inversions in 85% to 95% of all soundings. In their study, MERNILD & LISTON (2010) used vertical hand-carried temperature observations in the same study area to estimate the inversion frequency during summer (June-August in both 2005 and 2006). They also found a temperature increase with altitude in 84 % of the profiles. In contrast to the present study and that by GILSON (2018), only horizontal temperature profiles were used in the study by MERNILD & LISTON (2010).

No. of profiles	No. of profiles with inversions	Relative fraction	No. of SBI	No. of EI
113	94	83%	38	83

Table 3.8: Observed frequency of temperature inversions and inversion type (SBI, EI) in the UAV data.

Figure 3.13: Detected inversion types (EI blue, SBI red, EI + SBI green, no inversion no colour) at locations A, B, C, D, E, F, G and H by date and time of day. Measurements were made between 10:00 and 22:00.

Of these 121 observed inversions, there are 83 EI and only 38 SBI (Table 3.8, Figure 3.13). This agrees well with the findings from other comparable studies such as GILSON et al. (2018), GILSON (2018), and SERREZE et al. (1991). One reason for the high number of EI detected can be found in the nature of the respective inversion types. In contrast to SBI, EI can occur more than once in a profile. Another reason for the high number of EI is the proximity of the study region to the cold open water of the sea. This favours the advection of cold air from the sea further inland.

Figure 3.14: Wind speed and direction at all 4 AWS in the study area. The sea breeze can be seen, in particular in the AWS COAST data and somewhat weaker at AWS 196 as indicated by the dominant wind directions of S and SW and the increasing wind speeds during the day with a peak in the afternoon.

Such advection could be generated by local temperature differences between land and sea (sea breeze) or large-scale (synoptic) winds. The cold air from the sea is denser and pushes up the warmer air above the land causing a temperature inversion. There are several other studies that have observed a higher EI frequency along coastal regions (SERREZA 1991; FELIKS 1993; LU & TURCO 1994; IACOBELLIS et al. 2010; CARRILLO et al. 2015). This advection of cold air from the sea towards the land can also be clearly seen in the AWS data. In the AWS near the coast (AWS COAST and 196), winds are observed mainly from the south and southwest and they peak in the afternoon (Figure 3.14). The wind direction, its daily development, and its periodicity have the characteristics of a sea breeze (MERNILD et al.

2008; GILSON et al. 2018). However, the inversion type frequency in the study region is characterized by spatial and temporal differences. A high number of EI are identified at locations A, B, C, and D, while a higher number of SBI are observed at locations E, F, and G (Figure 3.13). Figure 3.15 illustrates the vertical temperature profiles at locations A, B, and C. Although the profiles from location A were collected one or two hours before those from locations B and C, an almost identical temperature change occurring with altitude is observed in all three profiles. The profiles recorded at location C are only slightly warmer than at A and B due to C's distance from the open water and its sheltered position at the end of the Mittivakkat valley. All three locations are above similar surfaces (sand and gravel) as shown in Figure 2.7.

Figure 3.15: The temperature profiles at location A (black), and at locations B (red), and C (blue), for which data were collected at the same time (B and C). The inversion base and top are also shown in the respective profiles. The grey, light red and light blue shading show a fog layer, and the green profiles indicate the pressure differences between locations B and C (C minus B, negative values indicate higher pressure at B while positive values indicate higher pressure at C). Data in black show the profiles we collected on our way up the transect, light blue data represent the profiles collected on our way down. The time at which data at A were collected is shown in black with B and C in red. The observed cloud cover is shown in octa next to the date, in black for location A and red for locations B and C.

If we look at the temperature profiles from around midday recorded close to the ground (0-20m a.g.l.) as shown in Figure 3.15 (black headings), we observe in most of the profiles that temperature decreases with height. Daytime solar radiation heats the land surface of the Earth and the sandy or stony surface transfers the heat into the overlaying atmosphere through the boundary layer's turbulent motion. At around midday, the surface and the air above are considerably warmer than the air at higher elevations. That is why we observe a steep temperature decrease with altitude close to the ground (0-20m a.g.l.).

The same, but already weaker, pattern can be observed in the evening profiles at locations A, B, and C (Figure 3.15, blue headings). There are a few exceptions to this, notably

the profiles from locations B and C on 16.07.2019 at around 15:00 and one profile at location C collected on 13.07.2019 at around 21:00. On 13.07.2019, at the time of the ascent, location C was already in shadow and the surface was losing energy due to radiative cooling while locations A and B were still receiving energy from the sun. 16.07.2019 was a cloudy day (Figure 3.15) and the sea breeze gentle (Figure 3.16). Clouds reflect and absorb a high fraction of the incoming solar radiation and reduce the amount of solar energy received at the surface. Consequently, the surface receives less energy, but loses more due to radiative cooling. A temperature increase was observed close to the ground on 16.07.2019 because turbulent mixing of the atmosphere was inhibited by the weak sea breeze and the radiative cooling due to the cloudy conditions. However, this temperature increase was not observed at location A. A possible reason for this could be the time difference of 5 hours or the proximity of location A to the sea. The time difference could have led to greater heat loss at B and C and the proximity of A to the sea to a cooling (albeit weak) of the air layers near the ground.

Figure 3.16: Relationship between the intensity of the sea breeze and inversion characteristics. The upper part shows wind direction and speed at AWS COAST (red) and 196 (blue). The charts in the middle indicate the interpolated temperature profiles for the days under review. In the lower part, the inversion characteristics of the profiles collected on these days are shown as bars, where the end of the bar corresponds to the inversion top, the beginning to the inversion base and the coloring to the strength.

While the atmosphere close to the ground is determined mainly by what happens at the surface, the intensity of sea breezes and distance from the ocean control the structure higher up. Z_{Base} values are lower on days with lower wind speeds near the coast (AWS COAST and 196) (3.14 and Figure 3.16 e.g., 16.07.2019) and higher on windier days (Figure 3.14 and Figure 3.16, e.g., 09.07.2019).

On some days we can observe that the profiles created in the evening are more homogeneous and have higher Z_{Base} values than those generated around noon. On sunny, clear days (Figure 3.15, e.g., 08.07.2019), the sea breeze builds up in the afternoon due to greater thermal differences between land and sea (Figure 3.14). The depth of the cold, moist marine layer becomes larger and pushes the inversion layer further up. This difference can also be seen in the pressure profiles (Figure 3.15). For example, on 09.07.2019 at 11:00 we cannot yet see a strong pressure gradient near the ground, but at 20:30 on the same day there is a pronounced pressure gradient from B (high pressure) to C (low pressure). Greater wind speeds also lead to greater turbulent mixing and thus smaller temperature gradients.

Figure 3.17: The temperature profiles at locations D (red) and E (blue) and the inversion base and top found in the respective profiles. The light red and light blue shading shows a fog layer. The green profile indicates the pressure differences between locations D and E (E minus D, where minus values indicate higher pressure at D and plus values indicate higher pressure at E). Data in black show the profiles collected on the way up the transect; data in light blue show the profiles collected on the way down. The time at which the profiles were collected is shown in red. The observed cloud cover is shown in octa (in red) next to the date for locations D and E.

The Z_{Top} variability observed is partly in contrast with earlier findings from MERNILD et al. (2008), who found, using RS observations, that air temperature inversions in summer (2006) usually occur up to 300m a.s.l. around Tasiilaq. In contrast, the studies by GILSON et al. (2018) and GILSON (2018) observed inversion-top heights (Z_{Top}) greater than 300m a.s.l. near Tasiilaq, even when considering the same years as MERNILD et al. (2008). Our study also found a high variability of Z_{Top} in the study region. This high degree of Z_{Top} variability was not only observed in the UAV data, but also in the RS and reanalysis products (Figure 3.10).

As with profiles from locations B and C, profiles at locations D and E were collected simultaneously. The SBI and EI frequencies at locations D and E are very different (Figure 3.13, Figure 3.17). While we observe EI, but not SBI, at location D, we find a large number of both SBI and EI at location E. This difference between the locations is a direct result of the difference in surface types. Location D is located above sand and gravel in front of the Mitivakkat glacier. Location E is located above melting ice at the tongue of the same glacier. As mentioned in the introduction

(1.2), the air temperature rises with distance from the surface of the melting glacier and creates an almost permanent SBI above the glacier surface, which is why such a high SBI frequency is observed at location E. The temperature gradient near the surface at location D is similar in shape to those at locations B and C. They all occur above the same surface type (sand and gravel) and the lapse rate is controlled by the same processes such as diurnal solar radiation pattern and cloud cover.

Figure 3.18: Deviation between profiles D and E, where the temperature profile of location D was subtracted from that of location E (E minus D, minus values indicate higher temperatures at D and plus values indicate higher temperatures at E). Data in black show the profiles collected on the way up the transect, data in red are the profiles collected on the way down.

The atmospheric temperature profiles at locations D and E become almost identical above that part of the surface layer that is modified by the surface properties. Figure 3.18 illustrates the difference between D and E in the first 50m a.g.l. The data reveal the greatest differences in the first 10m a.g.l, but already by 20m to 30m a.g.l. both profiles are more uniform and show only minor differences. Hence the depth of the air layer affected by the surface type is on average 10m to 20m high.

At both locations, Z_{Base} and Z_{Top} for EI are still controlled by the intensity of the sea breeze and the topographic configuration. When near-surface air is carried to a mountain ridge by thermally driven winds, such as a sea breeze, the cold, moist air is forced to rise along the mountain slope. Depending on the intensity of the wind, this air layer can either push the inversion layer further up or even penetrate the inversion (HENNE et al. 2005; LU & TURCO 1994; SERAFIN et al. 2018). For this reason, we observe slightly higher EI Z_{Base} values at locations D and E than at location C.

The thermal differences between the cold glacier surface and the warmer ambient air leads to the formation of a small-scale wind, known as a katabatic wind, which is a characteristic meteorological feature on alpine glaciers (AYALA et al. 2015). This wind can be detected in the pressure data of the profiles in Figure 3.17 and the AWS data (Figure 3.14). The katabatic wind also has an influence on the characteristics of EI at location D. On days when the katabatic wind was particularly strong (e.g., 13.07.2019, 16.07.2019), the cold air flowed from the glacier to location D and lifted up the warmer air layers nearer the ground, forming an EI. The pressure at location E is greater than at location D up to about 10-20m a.g.l., which is exactly where the Z_{Base} of the EI at location D is located on these days (Figure 3.17).

Locations F and G, sited above melting ice or snow on the Mittivakkat glacier, are similar to location E. A high SBI frequency is also observed at those locations due to the glacier atmosphere (Figure 3.13). At all locations on the glacier, SBI ΔZ and ΔT vary spatially and temporally. For example, the thickness and temperature difference of the SBI on the glacier depends on the temperature of the air layers above the glacier. During the field campaign, the lowest temperatures were recorded between 09.07.2019 and 11.07.2019 and the highest between 13.07.2019 and 16.07.2019. During the cold period, the temperature difference between the glacier surface and the surrounding air was low and, consequently, the ΔT in the SBI were small. During the warm period, the gradient between the glacier surface and the surrounding air was larger and thus the ΔT in the SBI greater.

Although locations F and G are at higher elevations than location E, there were several days where parts of both were warmer compared to location E. For example, higher temperatures were observed at location F on 12.07.2019, 13.07.2019, and 16.07.2019 (Figure 3.19) and at G on 10.07.2019, 11.07.2019, and 18.07.2019 (Figure 3.19), due to low-lying temperature inversions above this part of the glacier.

Figure 3.19: Temperature profiles at location E (black), F (red), and G (blue), and the inversion base and top found in each profile. Data in black show the profiles collected on the way up the transect, data in red show the profiles collected on the way down. The times at which the profiles were collected are shown in black for location E, red for location F and blue for location G.

There are only a small number of profiles available from location H. The four profiles include one inversion, which is an EI. The location is exposed on a nunatak and strongly influenced by the nearby glacier. According to the study of MERNILD et al. (2008), this location is dominated by katabatic winds. The ongoing turbulent mixing at this location suppresses the development of inversion layers.

3.2.2 Profiles along the surface

If we look at the horizontal temperature and humidity structure at 12m a.g.l. along the transect and also at the transect's surface types, a clear spatial pattern emerges. Daily and mean 12m a.g.l. air temperatures and relative humidities from the UAV profiles above the locations are shown in Figure 3.20. The two locations above sand and gravel which are closest to the sea (A and B) display lower mean temperature values compared to the other locations above a similar surface (C, D and H).

Figure 3.20: Horizontal temperature and relative humidity profiles at 12m a.g.l. at the locations per day and on average. The information was taken from the UAV profiles at 12m a.g.l. The distance corresponds to the sum of the distance between the locations. The letters indicate the locations, with colour representing surface type, black: sand and gravel; red: snow and ice.

From location A and B to location C, air temperature increased on average by 2°C. The temperature gradient from the coast into the Mittivakkat valley is driven by the sea breeze, the nearby ocean acting as a near-surface cooling agent. The influence of the nearby ocean is also clear in the humidity structure between A, B, and C. Location A has the highest relative humidity values of all locations above sand and gravel due to its proximity to the ocean. The average humidity drop of 12% between locations A and B is probably influenced by the greater distance from the ocean at B.

As already mentioned, location C is sheltered and therefore is on average warmer and dryer than at other locations above the same surface. The temperature at position D is on average 1°C colder than at location C, but the humidity is around 9% higher. Both values

might be determined by either the sea breeze or the glacier atmosphere. There is a sizeable temperature drop of 4°C between location D above sand and gravel and position E above melting ice. The temperature at this low level remains on average higher than that at the locations on the glacier (F and G) with a slight decrease up-glacier of 1°C . Overall, the relative humidity at the locations above snow and ice is around 50% on average, apart from position E, which displays the highest humidity values, averaging up to 65%. As already mentioned, there are only four temperature and two humidity profiles from position H, which makes an accurate evaluation of the typical temperature and humidity structure above this position more difficult. What we see from the few profiles is an average temperature increase of 5°C between the locations above the glacier and location H above sand and gravel. The humidity also increases between the two different positions.

Figure 3.21: Horizontal profiles showing the temperature at about 1.50m a.s.l. along the path walked that day. The distance is the distance from location B to the corresponding point.

The same temperature and humidity trend along a similar transect in the Mittivakkat area was observed in the study by MERNILD et al. (2004). They carried out manual measurements of the temperature and humidity from Sermilik, through the Mittivakkat valley, to the glacier margin, and then up-glacier to 220m a.s.l. These findings indicate that the horizontal temperature and humidity structures above the different locations are mainly influenced by distance from the ocean, wind regime, local topography, and surface type.

The same near-surface temperature structure along the transect is seen in the daily horizontal profiles (Figure 3.21). These horizontal profiles were collected at approximately 1.50m a.g.l. using sensors mounted on our backpacks. As already mentioned, the temperature of the air layer from the coast along the Mittivakkat valley (locations B to C) depends on the surrounding temperature and the intensity of the sea breeze. Low temperatures and a strong sea breeze can lead to a cold layer of air reaching far into the Mittivakkat valley (08.07.2019 and 09.07.2019), while warm temperatures and a weak sea breeze can lead to a contrasting picture, with a warm layer of air in the Mittivakkat valley (16.07.2019). Temperature increases up the glacier could be observed, especially on 10.07.2019, 13.07.2019, and 16.07.2019. On 13.07.2019 measurements were made while we were climbing a nearby nunatak instead of following the usual transect, and in the profile it is easy to see how air temperature increases with altitude.

3.3 Impact of synoptic scale events on temperature inversions

In this section, we outline and discuss the effect of synoptic scale events on temperature inversion characteristics. In 3.3.1, we give an overview of the synoptic conditions during the classified weather situations. The events are classified based on field notes, meteorological variables from all 4 AWS, and the analysis of ERA5 synoptic weather charts. Altogether five different major weather situations were observed between 08.07.2019 and 18.07.2019. In 3.3.2, the relationship between the inversion activity and intensity (f , ΔZ , ΔT , Z_{Base} , Z_{Top}) and the prevailing weather are analysed.

3.3.1 Classified synoptic scale events

3.3.1.1 08.07.2019: Occlusion with cold front character

The 500hPa geopotential height indicates a low-pressure zone coming from Iceland, which reached Sermilik at 16:00 on 08.07.2019. The cold air masses from the trough are clearly seen moving closer towards Sermilik in the thickness map below the 500 hPa geopotential height (Figure 3.22). The AWS data (Figure 3.22) indicates that the leading edge of the advancing cold air mass arrived in Sermilik between the nights of 07.07.2019 and 08.07.2019. At this time (approximately 07.07.2019 22:00), an abrupt drop in temperature and increase in humidity were observed, first at COAST and then, approximately 7 hours later, at 196. This was accompanied at lower elevations by shifting wind and low wind speeds. The colder, dense air at the ground was pushing up the less dense warm air leading to an increase in temperature and decrease in humidity at around 02:00 on 08.07.2019 at the CCMORE and GEUS weather stations. As soon as the leading edge of the cold air arrived at higher altitudes, the same abrupt, but less strong, temperature drop and gusty winds were observed at CCMORE (14:00). However, the temperature remained warmer at CCMORE than at COAST throughout the whole day. The increasing wind speeds at COAST and 196 at approximately the same time as arrival of the leading cold air edge at higher altitudes could be the reason for the higher temperature at CCMORE, with the colder maritime air layer pushing the cold, but warmer, continental air layers upwards. Just a few hours later (starting at 18:00) a strong wind was observed at CCMORE.

These changes in air temperature and stratification are also reflected in the interpolations (Figure 3.23 a). The temperature difference between the layers is more pronounced in the interpolated profiles created between 10:00 and 17:00 than in those created between 18:00 and 22:00. It can also be seen how the warm air is increasingly replaced by cold air masses.

Figure 3.22: Meteorological data from all AWS measured during the field study (2.2). The water level data of the proglacial lake and river (2.4) are shown, with the river data showing the measured distance from the reference point to the water surface. The dashed lines represent the time period during which the profiles were collected up (black) and down (green) the transect. Also shown are the ERA5 synoptic charts with pressure and atmospheric thickness patterns typical of the respective weather class. The dates when a particular weather class was observed are colour coded: light blue for “Occlusion with cold front character” (3.3.1.1), grey for “Cold sector of a depression” (3.3.1.2), light green for “Ridge” (3.3.1.3), yellow for “Trough” (3.3.1.4), and light red for “Warm sector of a depression” (3.3.1.5.) class.

Figure 3.23: The interpolated temperature profiles are divided into those collected on the way up the transect and those collected on the way down. The letters indicate whether the interpolated profiles were collected during the weather class a) “Occlusion with cold front character”, b) “Cold sector of a depression”, c) “Ridge”, d) “Trough” and e) “Warm sector of a depression”. The red dashed line between 1742 and 2119 shows the approximate starting point of the glacier.

3.3.1.2 09-11.07.2019: Cold sector of a depression

The 500hPa geopotential height map (Figure 3.22) indicates that a low-pressure system with a height of 5,568m passed Sermilik on 08.07.2019 and remained to the north of Sermilik until 11.07.2019. The atmospheric thickness was low during this time, indicating cold air above Sermilik. These cold conditions can also be seen in our AWS data (Figure 3.22) and interpolations (Figure 3.23 b). This was the coldest period during the entire field campaign.

This period was also characterised by strong sea breezes at COAST and 196, with wind speeds of up to 5m/s. Especially on 09.07.2019, high wind speeds, cold temperatures, and high humidity could be observed at both AWS. This cold and strong sea breeze pushed up the warmer continental air and led to higher temperatures at higher elevations (CCMORE). This contrasts with the wind speeds at GEUS, which ranged from 0m/s to a maximum of 1m/s during the entire period. The wind speed at CCMORE was low on 09.07.2019 and 11.07.2019 (1m/s), but relatively high on 10.07.2019, with wind speeds of up to 5.5m/s.

During this period, temperatures slowly increased, first at COAST and 196 (09.07.2019 01:00) followed by GEUS and CCMORE (10.07.2019 00:00), but remained low. The incoming solar radiation was low at COAST and 196 during this period due to cloud cover (low or high clouds). 09.07.2019 was foggy and cloudy (6 to 7 octas). However, no fog was observed on 10.07.2019, but there were high 8 octas clouds until 15:00. 11.07.2019 was a calm and sunny day with high clouds (4 to 8 octas) most of the day.

The interpolated profiles (Figure 3.23 b) show coolness of the air in this weather class in contrast to the other weather classes. In the interpolated profiles of 09.07.2019 between 10:00 and 16:30, the cold air is seen penetrating far into the Mittivakkat valley due to the strong sea breeze. The slowly increasing air temperature over the course of the three days is also visible.

3.3.1.3 13.07.2019: Ridge

Figure 3.22 shows the 500hPa geopotential height and atmospheric thickness pattern on 13.07.2019. An elongated area with a high 500hPa height and thickness (warm air masses) is seen moving from the south of Greenland towards Sermilik. This ridge reached Sermilik in the late afternoon of 13.07.2019.

The AWS data (Figure 3.22) display a sharp rise in temperature and decreasing relative humidity at CCMORE from approximately 16:00 until 17:00. A similar, but not so intense, temperature rise is seen at the other AWS. The lowest temperature and highest humidity on this day was measured at COAST. The wind speed at this location was moderate for most of the day (2m/s-3m/s). The temperature at 196 was higher, humidity lower, and the wind speed low compared to COAST, indicating a weak to moderate sea breeze bringing cold air from the sea further inland, but only at low elevations (COAST). The warm, less dense air from the ridge slid over the cold air masses from the sea and created a mixed unstable layer topped by a stable layer (Figure 3.23 c). At around 20:00 the wind became more intense at higher elevations (CCMORE), while in the lower area it became calmer (COAST, 196 and GEUS). 13.07.2019 was cloudy (up to 8 octas) and foggy in the morning. In the afternoon the sky cleared and the weather was sunny for the rest of the day.

3.3.1.4 15.07.2019: Trough

On 14.07.2019, a trough (an elongated area of relatively low atmospheric pressure) built up south of Iceland. The already well-developed high-pressure system to the east of the trough was still bringing warm air towards Sermilik. According to the 500hPa geopotential height map, the trough approached Sermilik at around noon on 15.07.2019 (Figure 3.22). The start of drizzle at approximately the same time was recorded in our field notes, but this was not reflected in the AWS data. The atmospheric thickness pattern indicates moderate temperate air masses above Sermilik.

CCMORE stopped logging data at around 01:00 on 15.07.2019. On this day there was a sharp temperature increase and decrease in humidity at COAST and 196 (Figure 3.22). The wind was gusty at both stations during the whole day. The highest wind speeds were observed at COAST and 196 at 03:00 (untypically from E to W) and 17:00 (4m/s from S to N), while the lowest wind speeds (0m/s-2m/s) were between those times, a pattern that does not follow the observed diurnal wind speed pattern at these AWS. The temperature at COAST followed the same pattern: high temperatures at 03:00 and 17:00 accompanied by high wind speeds and decreasing (but still high) temperatures accompanied by low wind speeds. Overall, the temperature at GEUS decreased moderately over the day. The wind blew mainly from E to W and showed the highest wind speed measured at this AWS during the whole field study at 01:00 and remained at 2m/s, a high level for this station. The incoming radiation on this day was the lowest recorded during the field campaign. The whole day was cloudy (8 octas) with drizzle and strong katabatic winds on the glacier.

3.3.1.5 16.07.2019: Warm sector of a depression

On 16.07.2019, the trough developed into a closed low-pressure zone (Figure 3.22). The Sermilik area remained in the outermost area of the low during the day. The thickness map indicates that an area of warm air was moving from the NE towards Sermilik.

During 16.07.2019, the temperature at COAST and 196 increased (after a sharp drop at around 01:00) and the humidity decreased (Figure 3.22). Over the whole day 196 showed higher temperature values than COAST and the wind speeds at both AWS were the lowest during the field campaign (0m/s-2m/s). The wind was gusty at both stations during the whole day. Over the day, the temperature at GEUS increased and the humidity decreased. The highest temperatures were at around 22:00. The wind speed was moderately high for this station (3m/s) and from E to W, which indicates a strong katabatic wind. The incoming solar radiation was still low. There were high clouds for most of the day (6 to 8 octas before 18:00 and 1 octa afterwards) and sunny conditions in the afternoon. The interpolated profiles (Figure 3.23 e) show clearly how much warmer it was on this day compared to the other weather classes, with the temperature also high above the glacier.

3.3.2 Effect of synoptic scale events on inversion activity and intensity

Inversion activity refers to the ratio between the number of profiles in which EI were found and the total number of profiles obtained during a particular time period. This ratio is expressed as a percentage (fraction of occurrence). The time periods include the profiles collected daily. They are further subdivided into the profiles collected on the way up the transect and those collected on the way down during the respective day. The inversion activity only refers to the ratio of profiles with EI to the total number of profiles. SBIs were not included because they occurred on the glacier every day, regardless of the weather conditions, due to the melting glacier surface. SBIs were found exclusively on the glacier with the exception of their occurrence in three profiles.

The intensity describes the thickness and strength of the inversion. To make the effects of weather events on the inversions clearer, the inversion characteristics (Z_{Base} , Z_{Top} , ΔT , ΔZ) found in the profiles are subtracted from the median of the respective inversion characteristics.

Occlusion with cold front character (OCF)

Figure 3.24 shows that EI were frequent during the OCF class, with EI occurring in all profiles collected on that day (08.07.2019). Similarly, most EI were thicker (ΔZ) and had more strength (ΔT) than the median until late afternoon. The upper (Z_{Top}) and lower (Z_{Base}) EI limits increased throughout the day, leading to a decrease in both thickness and strength. The EI from the last ascent on this day is particularly different from the others, as both thickness and strength and inversion top and base are below the median. In this weather class, the thickness, and the inversion top of the SBI was around the median and the strength was mostly below the median. The strength decreased with increasing time of the day and increasing altitude on the glacier.

As discussed earlier, during this synoptic event cold air was advected to Sermilik and the warm air ahead of the cold air was pushed up (Figure 3.23 a). The stable stratification (Figure 3.26 a) created by the advection of cold air remained after the leading edge of the cold air passed, due to the onset of a strong cold sea breeze afterwards. This led to a large number of EI during this day (100%). The low wind speeds at lower elevations (COAST and 196) at the onset of the cold front favoured the development of thick and strength EI. The sea breeze, which increased during the day, reached its peak in the late afternoon. The stronger the sea breeze became, the higher the EI were lifted, which is why the inversion base and top rose during the day. The last ascent of the day was made at a time when the sea breeze had just peaked, and wind speeds were increasing at CCMORE. The air was already well mixed and cool at this point. For this reason, the EI observed at this time is weak, the inversion base and top very high and not very strong.

Figure 3.24: The crosses show the frequency in percent, with the first cross per day representing the profiles collected on the way up the transect and the second for those collected on the way down. The inversion characteristics have been subtracted from their respective median values. 0 represents the median. Values above this are larger than the median, those below smaller. The inversion characteristics of the SBI are shown in red and those of the EI in blue. The individual weather classes are highlighted in different colours: light blue for “Occlusion with cold front character” (3.3.1.1), grey for “Cold sector of a depression” (3.3.1.2), light green for “Ridge” (3.3.1.3), yellow for “Trough” (3.3.1.4), and light red for “Warm sector of a depression” (3.3.1.5.) class.

Cold sector of a depression (CD)

EI were present in all profiles of the CD class surveyed between 10:00 and 16:30 on 09.07.2019, but by the evening (18:00-22:00), EI were found in only 70% of the profiles (Figure 3.24). This number decreased again on 10.07.2019 to 40% of the profiles collected between 10:00 and 15:30, then in the evening (16:00 - 22:00) increased again to 75%. On the last day in this weather class (11.07.2019), there were EI in only 50% of the profiles surveyed between 10:00 and 15:30, and, as on previous days, the proportion of EI in the evening (17:30-22:00) increased to 80%.

The EI thickness at location A was still below the median on the morning of 09.07.2019. As the day progressed, the thickness increased slightly and was at, or slightly above, the median. In the late afternoon to evening, the EI thickness fell sharply below the median again. EI strength was at or above the median in the morning and slightly below in the afternoon to late evening. The EI top and base were low at the beginning of the day, but rose strongly throughout the day.

Figure 3.25: UAV temperature profiles by day and weather class (a, b, c, d, e) are shown, with only the values at 0m, 100m, 200m, 300m, 400m, and 500m connected. The 100m segments (2.5.) are colour coded according to the LR calculated for them: blue (unstable), green (neutral), or red (stable). The numbers in red next to the profiles show the LR calculated for the respective segment.

SBI thickness was at, or very slightly below, the median. SBI strength was strongly below the median. In fact, these were the lowest values measured during the field campaign. On 10.07.2019, all the inversion characteristics of almost all the profiles were at or below the median. The top and base of the EI were rising, albeit at a low level. During 11.07.2019, the EI became thicker and were well above the median in the afternoon. The strength remained around the median throughout the day. The top and base of the EI were well above the median for most of the day.

Figure 3.26: Interpolation of the calculated potential temperatures in the profiles, divided into those collected on the way up the transect and those collected on the way down. The letter indicates whether the interpolated profiles were collected during the weather class a) “Occlusion with cold front character”, b) “Cold sector of a depression”, c) “Ridge”, d) “Trough” or “Warm sector of a depression”. The red dashed line between 1742 and 2119 shows the approximate starting point of the glacier.

The strong sea breeze at COAST and 196 and low wind speed at higher elevations led to a high number of EI in the profiles collected between 10:00 and 16:30 on 09.07.2019 (100%). The cold, moist maritime layer was deep and lifted the base and top of the inversions. Throughout the day, the sea breeze increased and led to a thicker, cold, well-mixed layer near the ground (Figure 3.26 b). Figure 3.25 b shows how the 100m LR changed throughout the day in the profiles near the coast (A, B, C, D) due to increasing wind speeds. The first 100m in the profiles were labile throughout the day due to the sea breeze. Above this unstable

layer, the stratifications in the profiles were neutral to stable between 10:00 and 12:00. As wind speed increased, the cold, well-mixed layer near the ground became thicker and led to unstable to neutral stratification in the first 300m to 400m in the profiles collected between 19:00 and 22:00. The deep maritime layer reached far above the glacier, suppressing SBI thickness and strength.

The deepening of the low-pressure zone on 10.07.2019 led to higher wind speeds at all elevation levels (except GEUS). This encouraged the development of a mixed, unstable lower planetary boundary layer in the study region. The 100m LR on this day (apart from the first 100m in the profiles at the glacier) was labile to neutral in all profiles, indicating a well-mixed low atmosphere (Figure 3.25 b). Inversions are less likely to occur during such strong wind events (MERNILD & LISTON 2010), which is why the number of EI was small and the strength and thickness were below the median value.

Lower wind speeds were observed at higher elevations and higher wind speeds at lower elevations during 11.07.2019. The wind speeds at CCMORE and 196 were particularly high at the start of the measurements and decreased slowly thereafter. The thick maritime layer is also reflected in the 100m LR of profiles A, B, and C measured between 10:00 and 11:00. The layer is very unstable in the first 100m and neutral above (Figure 3.25 b). Colder air from the sea flowed inland and lifted the warmer air layers on the coast. This led to a stable stratification of the lower atmosphere and a high number of inversions in the profiles (80%) collected in the late afternoon to noon. In general, the SBI thickness and strength on 11.07.2019 were at or below median values, except for location F in the afternoon. Here the SBI top was well above the median. The reason for this is probably a low-lying inversion layer above the glacier. This low-lying inversion layer led to an extension of the SBI already present at location F.

Ridge (RG)

During the entire RG class, one or more EI were found in every profile (100%) (Figure 3.24). The thickness and strength of the EI were sometimes well above the median. The EI top was above the median in the morning until noon, and the EI base at or below median. In the afternoon to evening, the EI top and base were both far below the median. In the evening (20:30 - 22:00), the EI strength was high, but the thickness was less than in the profiles considered previously. For the SBI, all characteristics were well above the median on this day. The SBI of this weather class were the thickest and the strongest.

Warm air advection occurred ahead of an approaching ridge on 13.07.2019. The maritime layer was small because of the weak sea breeze. The warm air from the ridge rose gradually over the small, cold, maritime layer and produced a stable stratification (Figure 3.26 c). The contact of warm and cold air masses caused the high number of EI. Low to zero wind speeds at all elevations above the maritime layer led to thick, stable, and strong EI. The weak sea breeze meant that the warm air layers over the land were not lifted very high. Therefore,

this warmer air layer lay low and covered parts of the glacier (location E to location F) in our study area. This low-lying inversion layer led to an extension of the SBI already present at all locations on the glacier.

From 20:00, wind speed decreased at COAST and 196, while it increased at CCMORE. Due to the weaker sea breeze, the cold maritime layer became narrower and the warmer air layer above sank lower, resulting in lower EI lower limits. Stability and strength increased due to the ridge moving closer along with its warm air. At this time, the inversion layer was still above the glacier and led to thick SBI.

Figure 3.25 c shows how the 100m LR changed throughout the day. At the beginning of the measurements, the sea breeze was moderate and the first 100m to 200m of profiles A, B, and C (collected between 10:00 and 11:15) were unstable to neutral. The 200m to 300m above this layer were stable. In the afternoon to evening, as the wind speed increased at CCMORE and decreased at COAST and 196, the stable layer sank and the well-mixed neutral layer expanded above it.

Trough (TG)

In the period between 10:00 and 15:30 on 15.07.2019, EI were found in 80% of the collected profiles (Figure 3.24). The thickness was at, slightly below, or slightly above, the median with the inversions at locations A and B being thicker and lower than at the other locations. EI strength was at or below the median. SBI thickness was at the median, except for at location F, where it was well above. SBI strength was well above the median at all locations.

During the time when the profiles were being collected, the wind speed at CCMORE and 196 was low (0m/s-2m/s). For this reason, the cold maritime air layer and the warmer air layer above it were also low, which in turn led to low EI tops and bases. The air was warm on 15.07.2019, although slightly colder air masses from the trough had reached Sermilik from the early morning. The lower atmosphere was well mixed from the strong winds during the morning and the day before.

The 100m LR from this day shows that clearly. The stratifications are, with a few exceptions, unstable to neutral (Figure 3.25 d). The whole day was overcast. High cloud cover generally weakens temperature inversions in the Arctic due to enhanced downward long-wave radiation (KAHL 1990; SERREZE et al. 1992; HUDSON & BRANDT 2005; GILSON et al. 2018; NIELSEN-ENGLYST et al. 2019). For these reasons, thickness, stability, and strength were around the median and activity moderate. The weak, gusty winds at lower elevations led to a low-lying weak inversion layer. On this day, the inversion layer lay just above and on top of the glacier, also increasing the thickness of the SBI.

Warm sector of a depression (WD)

The number of EI found was moderate at the beginning of this weather class on 16.07.2019, with EI found in 60% of the profiles collected between 10:00 and 17:00 (Figure 3.24). The thickness and strength were below the median during this period. Two EI were often found in one profile, one low, the other higher. This is reflected in upper and lower limits that are far above or below the median. In the evening, between 18:00 and 22:00, only three profiles were collected and an EI was found in all of them. The profile at location A is particularly striking due to its top and base being far below the median, and its moderate thickness and its strength that is slightly above the median. The SBI strength was the highest of all the weather classes and the thickness was at the median except for location G in the late afternoon.

The wind decreased at all elevations, apart from GEUS, as soon as Sermilik was in the pressure zone. The major thermal differential between the surface of the glacier (0°C) and the warm surrounding air led to one of the strongest katabatic winds of the field campaign. The wind's strength meant it reached location D and lifted the warmer air layer and formed a small, weak EI. In the late afternoon, the katabatic wind reached its peak and lifted this weak EI a little more.

This intensive mixing of the lowest air layer at location D is also reflected in the 100m LR Figure 3.25 e (the only profile with an unstable stratification in the first 100m). Due to the SBI at locations B and C and the strong katabatic wind leading to a weak EI at location D, the upper and lower limits on this day were among the lowest in the study. In general, the lower atmosphere was neutrally stratified on this day, with few exceptions, and there was hardly any mixing. The absence of a sea breeze led to less thick EI during the whole day. Other studies have found that under calm, windless conditions the frequency of inversions in the Arctic is low, while at wind speeds around 3m/s-10m/s inversions are more frequent (HUDSON & BRANDT 2005; MILLER et al. 2013; GILSON 2018; NIELSEN-ENGLYST et al. 2019).

3.4 Impact of temperature inversion statistic on ablation

This section focuses on the possible impact of temperature inversion characteristics on the ablation of the Mittivakkat glacier. The absolute and relative proglacial lake and river level changes and the mass loss measured at the ablation stakes will be used to quantify the ablation variations during the field campaign and to link them to the inversion statistics. The ablation data is presented and discussed in the first subsection (3.4.1). Their interaction with the inversion statistics is presented and discussed in the second subsection (3.4.2).

3.4.1 Observed ablation variability

The daily absolute lake level (LL) fluctuation data in Figure 3.22 show an overall trend during the field study with maximum and minimum LL increasing from the beginning to the end of the observation period. This trend was interrupted between 14.07.2019 and 17.07.2019 with an overall increase of the minimum and decrease of the maximum LL. The data also show a diurnal cycle: significant LL increase from the morning (approximately 10:00) until the late afternoon (approximately 16:00) followed by a LL decrease until the morning of the next day.

Overall, the LL rises if the lake experiences a net inflow of water and falls if there is a net outflow of water. During the field study, this lake received input exclusively from the Mittivakkat glacier. (It is assumed that the short-lived drizzle on 13.07.2019 did not have a significant impact on the lake fluctuation.) Glacier (ice and snow) melt is governed by the atmospheric energy fluxes, which are majorly affected by incoming solar radiation, air temperature, and wind (SWIFT 2014).

Figure 3.22 clearly shows this relationship between the increase or decrease in incoming solar radiation and the increase or decrease in LL. The diurnal variation of the LL follows the diurnal variation of the incoming solar radiation with a time lag of a few hours that is caused by the time needed for the passage of the melt water through the glacier, snow and firn or temporally stored water in glacier pockets or soil (FOUNTAIN & WALDER 1998). The average lag is 3 to 4 hours.

The correlation between LL, water temperature (WT), and the meteorological variables from the 4 AWS was calculated using the Pearson correlation coefficient (r). Figure 3.27 shows the correlation between the LL, WT, and the meteorological variables measured: at the same time (-0H), 1 (-1H), 2 (-2H), 3 (-3H), 4 (-4H), 5 (-5H), and 6 (-6H) hours before. -4H shows a strong relationship between LL and incoming solar radiation. The relationship is particularly strong between GEUS and LL ($r = 0.84$). A moderate to strong relationship can be observed between LL and temperature at all AWS, except GEUS.

Figure 3.27 also shows a weak negative correlation between LL and humidity at COAST and GEUS and a moderate negative correlation at 196 and CCMORE. This indicates that

the higher the humidity, the lower the melt. One reason for this could be colder and more humid air advected from the sea towards the land pushing itself under the warmer and drier continental air layer (formation of EI). Depending on the temperature, the air can hold different concentrations of moisture (dew point temperature) before it condenses and forms fog. On days with particularly low temperatures, such as 08.07.2019, 09.07.2019, 11.07.2019 and 12.09.2019, we therefore also observe fog under the inversion layer. Fog reduces the incoming solar radiation, which in turn reduces the melting.

Figure 3.27: Pearson correlation coefficients calculated between water temperature (WT), water level fluctuations of the proglacial lake (LL), and the meteorological variables of the 4 AWS. The correlations are calculated between simultaneously measured meteorological variables: short-wave radiation (SW), air temperature (AT), wind speed (WS), and relative humidity (RH). Temporally shifted values for WT and LL are also given for 1, 2, 3, 4, 5 and 6 hours before the measured WT and LL values in order to detect temporally delayed cause and effect processes.

In contrast to the LL, the WT has a stronger relationship with, and reacts more directly to, the amount of incoming solar radiation (Figure 3.27 WT). This is because, after heat transfer by absorption of solar radiation and consequent heat conduction within the water column, there is no further delay due to transport (NIEDRIST et al. 2018).

The 09.07.2019 was foggy at lower altitudes (COAST and 196) and sunny at higher altitudes. The fog line extended to just below the AWS GEUS until late afternoon, after which the fog also rose to higher altitudes. Above the fog, there were high clouds of 6-7 octas until early afternoon (14:00), of 2-3 octas from midday until evening (20:00) and of 7-8 octas afterwards. There was a similar diurnal cloudiness pattern on 10.07.2019, with high clouds until 15:00 (8 octas) followed by sunny conditions with only a few clouds (2 octas) until the evening. 11.07.2019 was partly cloudy to partly sunny throughout the day (4-7 octas). During these days, which were partly cloudy but also partly sunny, there was a moderate relative LL increase of about 5cm every day and a similar diurnal LL pattern (Figure 3.28).

12.07.2019 and 13.07.2019 were sunny with little cloud. This is reflected in the incoming solar radiation measured at the AWS. Figure 3.28 shows that on 12.07.2019 and 13.07.2019 a higher LL was reached earlier than on the previous days. The water temperature also reached higher values.

Figure 3.28: Hourly meteorological data from AWS GEUS, which is located directly on the glacier. The relative LL change by hour and by WT are shown in the bottom chart.

The period with the lowest incoming solar radiation was between the afternoon of 14.07.2019 and 17.07.2019. The incoming solar radiation was reduced on 14.07.2019 at lower elevations (COAST, 196) and between 15.07.2019 and 17.07.2019 at all elevations (COAST to GEUS) (Figure 3.22). Already by the afternoon of 14.07.2019, the sky was cloudy (8 octas) 15.07.2019 was foggy and cloudy (8 octas) all day, 16.07.2019 and 17.07.2019 were overcast for half of the day, 16.07.2019 until noon and again in the afternoon, and 17.07.2019 until late afternoon. The dense cloud cover in the afternoon of 14.07.2019 is reflected in a lower maximum relative LL. The high cloud cover on 15.07.2019, which lasted all day, led to less melting despite the high air temperature, which in turn led to a lower maximum LL (the lowest value in the field campaign). It was not until 16.07.2019 that the increasing incoming solar radiation due to the less dense cloud cover led to a higher maximum relative LL again.

During the nights of 14.07.2019 to 15.07.2019 and 15.07.2019 to 16.07.2019, the LL remained at a higher level than in the days before. This melting, which also occurred overnight, could have been caused by the heavy cloud cover during this period. Cloudiness warms the surface by increasing the incoming long-wave radiation, while the associated decrease in shortwave incoming radiation cools the surface. The direction and strength of the surface radiation budget modification from the cloud radiative forcing depends on the cloud properties (e.g., cloud phase, optical depth, elevation, temperature) and cloud cover (GILSON 2018; AMBACH 1974; SHUPE & INTRIERI 2003; TRICHT et al. 2015; CURRY et al. 2000). Low-level liquid clouds tend to increase surface warming in the Arctic due to increased long-wave radiation to the surface. This modification occurs during the middle of the arctic melt season over areas with low albedo when the solar elevation angle is at its highest (short-wave cooling) (GILSON 2018; SHUPE & INTRIERI 2003; BENNARTZ et al. 2013; WANG et al. 2016).

This means that there is also an energy input into the snow and ice cover at night and thus more melting can take place at night on these days than on the other days. An additional reason for the high meltwater input in the night between 14.07.2019 and 15.07.2019 may be the strong katabatic wind during this time (Figure 3.28). Katabatic winds transport sensible heat to the snow and glacier surface through turbulence, thus enhancing melting (OHATA 1989; HOCK 2005).

16.07.2019 and 17.07.2019 were less cloudy and the air temperatures were high. The highest WT and LL were observed during these days. The katabatic wind was also strong on 16.07.2019 as was therefore the sensible heat flux to the glacier.

The daily proglacial river level (RL) variability depends on the glacier melt. The melt runoff into the lake peaks in the afternoon, as does the river level. Lower water levels in the river can be seen between 08.07.2019 and 13.07.2019 and higher between 15.07.2019 and 18.07.2019 (no measurement was taken on 14.07.2019). The highest water level in the proglacial river was observed on 16.07.2019 where it was 2cm above the measurement mark.

Figure 3.29 illustrates the daily ablation measurements and cumulative ablation at 5 stakes at different elevations on the glacier (2.4). At M01, a constant ablation of 4cm to 6cm per day was observed. Ablation stake M02 shows ablation fluctuations of 2cm to 6cm, with the lowest values measured on 10.07.2019 and 11.07.2019. At M03, the lowest ablation was measured on 11.07.2019, but there was very high ablation on 10.07.2019, and a high cumulative ablation on the 15.07.2019 (cumulative from 3 days), and 17.07.2019 (cumulative from 2 days). M04 also shows low ablation values on 09.07.2019, 10.07.2019 and 11.07.2019, but very high on 15.07.2019 which is the sum of the ablation from 12.07.2019 to 15.07.2019. However, even if the ablation values from M02 and M04 measured on 15.07.2019 are divided by the number of days for which the ablation stands (M02 2 days, M04 4 days), the value would still be higher than the previous daily values.

Figure 3.29: The upper chart shows the measured ablation at ablation stakes 1 to 5. The bottom chart shows the cumulative ablation over the study period. The red lines pointing to a bar indicate when the measurement started at the respective ablation stake.

In general, the highest ablation was measured at the lowest elevation ablation level M01 (Figure 3.29), but there were some days when more melt was measured at M03, which is located at 355m a.s.l., than at the lower ablation levels M01 and M02. On 09.07.2019, more melt is measured at M02 and M03 than at M01. A low-lying layer of fog in the afternoon to late evening on 08.07.2019 may have led to reduced incoming solar radiation at M01, while the ablation levels at higher elevations were exposed to solar radiation, resulting in more melting at higher than at lower elevations. The same effect could be the reason for the increased melt at M03 compared to M02 and M01 on 10.07.2019. A layer of fog over larger areas of the glacier was observed on 09.07.2019. At the time of measurement, M02 (09.07.2019 13:50) was in fog, while M03 was already above it. When this spot was passed again 4 hours later, M03 was still above while M02 and M01 were still below the fog line. On 10.07.2019 the sky was very cloudy, but no fog was observed, but the reduced incident solar radiation due to the fog cover in the lower parts of the glacier on the previous day could have led to the lower ablation values at M01 and M02 compared to M03. The sudden increase in ablation at M04 between 12.07.2019 and 15.07.2019 could be due to the advection of warm air masses from 13.07.2017 to 16.07.2019. In particular, on 13.07.2019 (Figure 3.22) we see how the air temperature increases with increasing altitude and thus higher elevations at the glacier are surrounded by warmer air than lower elevations are.

3.4.2 Interrelations

Figure 3.31 shows that in the period in which more melt is observed at the ablation stakes (13.07.2019 to 16.07.2019) the strength and thickness of the SBI on the glacier is higher than in the period with less melt (08.07.2019 to the afternoon of 11.07.2019). The strength of the SBI over the glacier is closely related to the surrounding air temperature. The glacier surface is at the melting point of 0°C during the ablation period in summer. This means that the strength of the SBI on the glacier is determined by the air layer above the glacier: the higher the temperature of this air layer, the higher the SBI strength. Figure 3.23 shows the different temperatures of the air masses during the field campaign. Here we see that between 08.07.2019 and the afternoon of 11.07.2019 the air temperature was colder than between 13.07.2019 and 16.07.2019.

The temperature difference between the glacier surface and the surrounding air together with the wind speed determine the strength of the turbulent sensible heat flux to the glacier. The higher the temperature gradient (the amount by which the air above the glacier is warmer than the air at the surface) and the higher the wind speed, the more sensible heat is transferred to the glacier and the more melting can occur. Furthermore, the higher the thermal difference between the glacier surface and the ambient air, the more intense the katabatic wind developed on the glacier. This is why the most intense katabatic winds were observed during the field study from 14.07.2019 to 16.07.2019 (Figure 3.31). Such katabatic winds enhance the sensible heat flux to the glacier as wind speed increases (OERLEMANS & GRISOGONO 2002; WANG et al. 2019; MOTT et al. 2020).

Figure 3.30: The approximate position of the glacier tongue in 1933 (red dashed line) compared to 2019. Basemap: Sentinel-2 2019; Digital elevation model: ArcticDEM; Data processing in QGIS.

The study by HASHOLT et al. (2016), which was also carried out in our study area, compared historical ablation measurements at the terminus of Mittivakkat glacier from 1933 with more recent measurements (1996-2012). The measurements taken in 1933 were made during

Figure 3.31: The daily inversion statistics with the profiles divided into those collected on the way up in the transect (in black) and those collected on the way down (in red). The dashed lines show the location at which a profile was collected. If an inversion was found in the profile, it is shown as a bar, with the bottom of the bar representing the inversion base and the top representing the top and the colouring representing the severity. The black line shows a rough indication of the topography along the transect.

one of the warmest East Greenland summers in the entire twentieth century. The study assumes that temperatures were the same during both ablation measurement periods. Nevertheless, the researchers found that ablation rates in the more recent period are higher at the termini than the historical readings. They explained the increase in ablation rates since 1933 by a reduction of surface albedo and the retreat of the Mittivakkat glacier out of the cold maritime layer. This assumption is supported by our observations. In the area which the glacier used to cover (Figure 3.30), we only observed a maritime layer narrower than 100m only on a few days (13.07.2019, 15.07.2019 and 16.07.2019) (Figure 3.31). From our data, it is difficult to determine how often the cold and humid air flowed from the sea to the present glacier tongue during our observation period. The pressure data from Figure 3.17 do not reveal this, but if we look at the AWS data we can see that on many days the sea breeze reaches as far as weather station 169 (e.g. 08.07.2019-11.07.2019) and on 10.07.2019, intensified by the large-scale weather situation when an air flow from the sea towards the glacier took place at all weather stations (Figure 3.22). However, in 1933 the glacier tongue reached almost to the end of the Mittivakkat valley and was about 143m lower than today (Figure 3.30) (today: approximately 150m a.s.l.) (HASHOLT et al. 2016). This means that in 1933 the glacier tongue was more frequently protected by the cold maritime air layer than today.

On days when the sea breeze reaches AWS 169, we also observe less melting at the ablation stakes. The EIs were also higher on these days than on days with weak sea breezes. Conversely, this means that an intense sea breeze and the resulting high inversion layer (high EI Z_{Base}) have a positive effect on the mass balance of the glacier (less melt). For example, the sea breeze was intense from 08.07.2019 to 11.07.2019. The EI Z_{Base} are high, as can be seen in Figure 3.31. The effects on the LL fluctuations are not so clear here. The LL reached a somewhat lower maximum from 09.07.2019 to 10.07.2019 and from 11.07.2019 to 12.07.2019 than on most days before and after, but from 10.07.2019 to 11.07.2019 the LL maximum was at a similar level to the other days.

In the period between 13.07.2019 and 16.07.2019 a less intense sea breeze was observed, with wind speeds being especially low on 15.07.2019 and 16.07.2019. Figure 3.31 shows that during this period the EI Z_{Base} are low and the SBI thickness and strength were high. We do not have ablation stake measurements from 14.07.2019 with which we could determine the amount of ablation from 13.07.2019 to 14.07.2019, but the measurements from 15.07.2019 to 16.07.2019 show higher ablation than on the previous measurement days. The LL shows a higher and longer maximum from 13.07.2019 to 14.07.2019, at which point the melting and thus the maximum and minimum LL decrease.

However, the effects and impact of the inversion characteristics on the ablation behaviour of the Mittivakkat glacier cannot be quantified so clearly on the basis of our data. What is clear, however, is that, among atmospheric variables studied, we find a strong correlation between LL and incoming solar radiation when applying a time lag of 3-4 hours, and with LL and air temperature when applying a time lag of 0 hours.

CONCLUSIONS AND OUTLOOK

The spatio-temporal variations and characteristics of temperature inversions in Ammassalik have so far mainly been determined using RS or reanalysis data. However, these data sets have several limiting properties such as low spatio-temporal resolution which result in a not entirely realistic picture of the temperature inversion variations in the region being shown. For a more fully picture of the temperature inversion characteristics in Ammassalik, UAVs were used to collect high temporal resolution profiles of atmospheric variables such as air temperature, humidity and pressure up to 500m a.s.l. during a 13-day field campaign in summer 2019. An algorithm was then used to quantify the temperature inversion characteristics such as type (EI or SBI), thickness, height, strength, and frequency in all profiles.

Using this high-resolution data, we quantified how the properties of RS and two commonly used reanalysis datasets, ERA5 and ERA-I, affect the detection of inversions and how well the temperature profiles of these datasets match those of the UAV. Among all the evaluated data sets, the ERA5 profiles have the strongest correlation with the UAV profiles and the lowest deviation while the ERA-I data have the lowest correlation and highest deviation. Temperature biases were found in the near-surface areas of all data sets evaluated. A cold bias was found at the surface in the RS and ERA-I data. This cold bias found in ERA-I could be due to differences between the topography used in the reanalysis model and the actual topography. The ERA5 data generally show a warm bias in the first 100m a.g.l., except for the level 2m a.g.l., where mostly colder temperatures are measured than in the UAV profiles. The correlation between all UAV profiles and all evaluated data sets is higher and the RMSE lower between the PM data. The ability of the evaluated datasets to reproduce realistic temperature inversion characteristics for the first 500m a.g.l. depends to a large extent on their vertical resolution and temperature biases. The high SBI number in RS and ERA-I, in contrast to the UAV data, is caused by the cold bias, while the high EI number in the ERA5 data is caused by the cold bias at 2m a.g.l. followed by a warm bias. The low resolution of RS and ERA-I also leads to an overestimation of inversion thickness and strength and an underestimation of EI. The better vertical resolution in the newer reanalysis product ERA5

allows for a better representation of the inversion characteristics.

The 113 high-resolution UAV profiles also made it possible to describe the influence of different surface types and the environment (e.g., proximity to the sea) on inversion formation and characteristics. It was possible to find inversions in 83% of all profiles. The most common type of inversion was EI (63%), the main reason being the proximity to the sea and the regular advection of cold air from the sea towards the land (sea breeze). The stony and sandy surface between the sea and the glacier warms up during the day and, together with the sea breeze, leads to the frequent formation of EI, while the surface on the glacier (snow and ice) acts as near-surface cooling agent and leads to the permanent formation of SBI. Near-surface lapse rates above sand and gravel are mainly governed by the insolation reaching the surface. On sunny and clear days, these surfaces warm up and transfer heat to the layers of air above them, leading to a decrease in temperature with height above this surface. The SBIs which are rarely observed over these surfaces (during the field campaign), form when there is more long-wave radiation emitted from the surface than short-wave radiation reaching the surface. While the atmosphere close to the ground is mainly governed by the surface, the sea breeze intensity and distance from the ocean determines the structure higher up. The stronger the sea breeze the higher the EI: with a moderate sea breeze the inversions are lower and thicker with no or very weak sea breeze the inversions are very low and thin.

The quantified inversion characteristics of the 113 UAV profiles were also used to determine the effects on inversion frequency (fraction of occurrence) and intensity (thickness, strength) of four synoptic scale events observed during the field campaign. The temperature inversions were most frequent and intense when two air masses of different temperatures met and the sea breeze was moderate. This was most pronounced on 08.07.2019 and 13.07.2019. On 08.07.2019, during an occlusion with cold front character, cold air masses moved towards Sermilik and pushed under the warmer air layer, while on 13.07.2019 warm air advection took place in front of an approaching ridge and moved above the narrow maritime air layer resulting from the weak sea breeze. The lowest frequency and intensity were observed when Sermilik was under the influence of a depression between 09.07.2019 and 11.07.2019. The strong winds and cloudiness observed during this period inhibited the formation of EI.

The final objective of this master thesis dealt with the effect of quantified inversion characteristics on the ablation of the nearby Mittivakkat glacier. Overall, the incident solar radiation could be identified as one of the most important driver of ablation during the field campaign while the contribution of the inversion layer to ablation could not be quantified exactly. The only observation was that on days when the EI layer was low and thick and had a high strength and the SBIs on the glacier were also thick with a high strength more melting occurred while on days when the EI layer was high and thin and had a low strength and SBIs on the glacier were thin with a low strength, the melting was lower. However, a complete answer to the question of how inversions affect ablation cannot be given due to the short study period and requires further study.

UAVs have proven to be a useful tool to perform high spatial and temporal resolution measurements of the lower atmosphere. In contrast to RS or reanalysis data, however, the achievable altitude is lower and the implementation more complex. Results from this study show, that UAV measurements could serve as a supplementary data source, especially to improve the vertical, temporal, and spatial resolution. This could lead to more accurate observations and research outcomes from studies of the state of the lower atmosphere. The temperature biases, especially in RS, should also be taken into account in subsequent studies and the causes investigated.

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