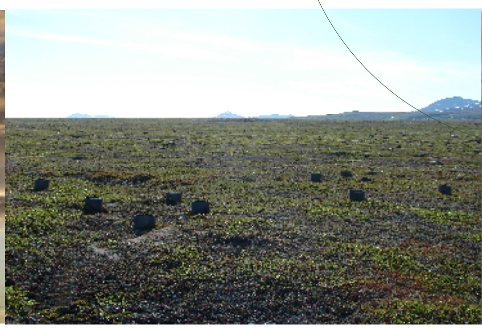
FACULTY OF SCIENCE
UNIVERSITY OF COPENHAGEN

Methane dynamics in a permafrost landscape at Disko Island, West Greenland

Field course 2011



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Foreword

The Field Course in Physical Geography 2011 took place between 1.-12. of July 2011 at Disko in West Greenland. In total 12 students participated in the course and have now completed the different parts: Preparations i.e., fine-tuning of projects, packing of equipment and the actual field course in Greenland. This report represents the final part for the course. The course was organized and lead by Birger Ulf Hansen and Bo Elberling from The Department of Geography and Geology.

The aim of the Field Course in Physical Geography 2011 was to integrate:

1. Meteorological gradients quantified within the landscape at Flakkerhuk and active layer thickness modeled over two thawing seasons based on soil thermal properties from local soil types

2. Active layer and permafrost characteristics based on samples collected from pits and permafrost coring

3. In-situ methane fluxes between the soil surface and the atmosphere using a mobile chamber setup

The combination of these three parts can provide new insight with respect to the current and future net methane budget within the study area taken current and future climate trends into account. The area Flakkerhuk at East Disko was chosen as the study area due to that the landscape is fairly well-described and due to the presence of wetland, which is considered hotspots for methane production.

Part of the course has been directly linked to PERMAGAS – a GEOCENTER Denmark project, which deals specifically with methane production and methane release to the atmosphere. Therefore, several other people took part in the field work at Flakkerhuk this summer including Thomas Friborg, Christian Juncker Jørgensen, Vibeke Ernstsen, Svend Funder and Niels J. Korsgaard.

A film project was linked to the field part of the course. For that reason, Pernille Semler and Bent Yde Jørgensen (Chilbal Film) participated in the field. During February 2012 a 30 min film will be presented regarding the activities in the field highlighting aspects of teaching and research "hand in hand". The film project is financed by Danish Agency for Science, Technology and Innovation.

Thanks to 12 enthusiastic students, the PERMAGAS project, Arctic Station (including Gitte Henriksen, Frantz Nielsen and Frederik Grønvold) as well as Faculty of Science, University of Copenhagen for financial supporting this Field Course in Physical Geography.

October 2011 Bo Elberling
(Course responsible)

Group picture

Group picture in front of Arctic Station

Back row:	Birger U. Hansen[1], Will Manning[2], Alejandro Barrera[3], Nicolai N. Christensen[4], Bo Elberling[5]
Mid row:	Toke Fogh[4], Anders Bjørnsson[4], Mette Hansen[4], Malene A. Hansen[4,6], Katrine M. L. Johansen[4], Sascha Müller[4]
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Introduction to the area of Disko and Flakkerhuk

In the following part an introduction to the field site (Figure 1) and the physical environment is given. Starting out, the geological development of Disko Island and processes forming the current landscape in Flakkerhuk is described. Then the local climate regime followed by a vegetation description is stressed. Subsequently, methane in terms of a potential greenhouse gas is explained and an overview of permafrost is presented. The relationship and interactions of methane and permafrost are put forward and leads to the aim of the field course expressed as each articles is shortly presented.

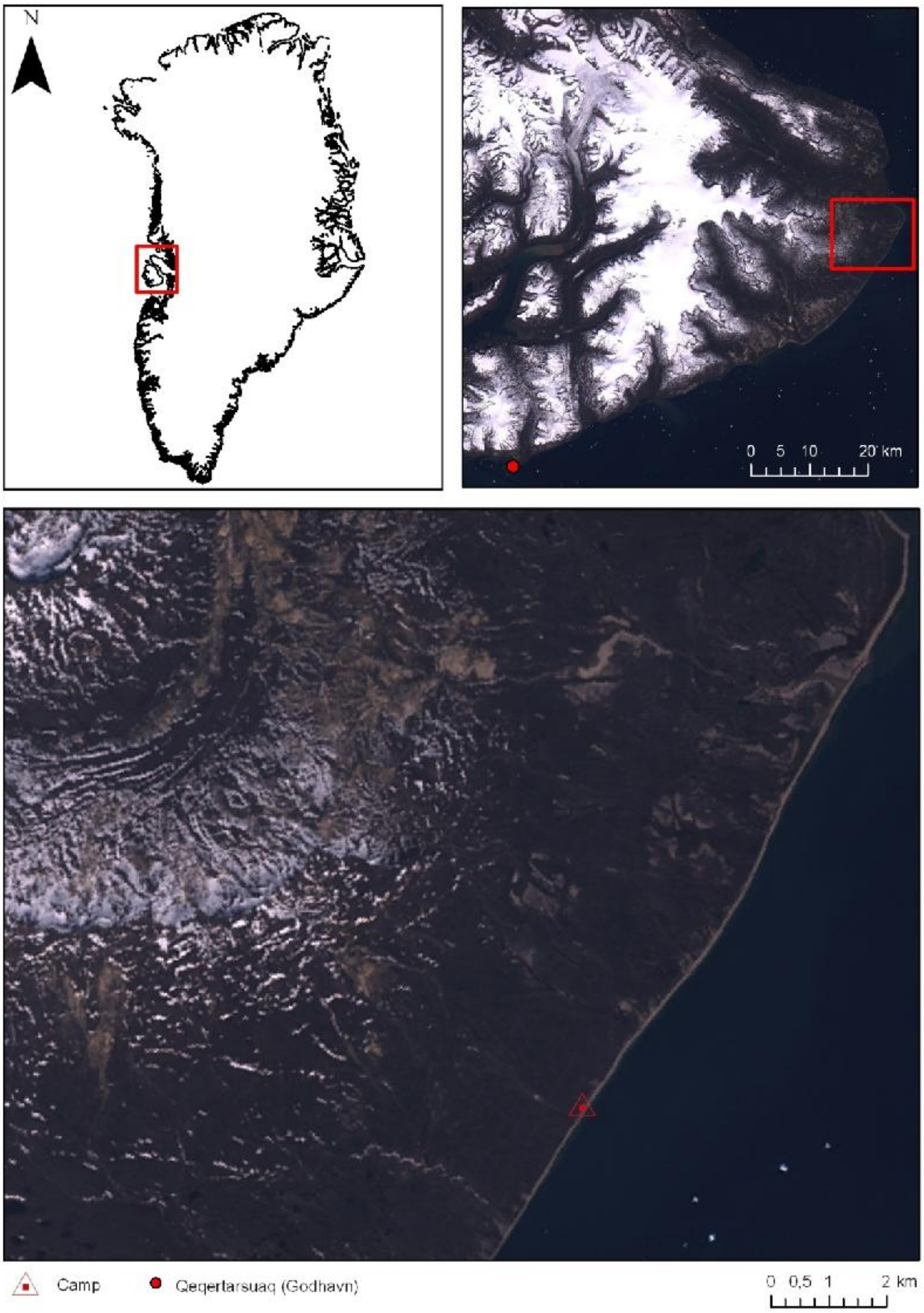

Figure 1: Overview of the field site at Flakkerhuk, Disko, West Greenland.

Geology

Disko has four major geological regions (Bruun et al., 2006); the archaic bedrock which is more than 1.600 million years old, tertiary sediments from the cretatious and the paleogenic era, with ages from 60 to 100 million years, volcanic rock formations from the paleogenic era with ages from 54 to 60 million years and sedimentary deposits from the quaternary period, which are less than 100.000 years old (Bruun et al., 2006).

In the time span from 1.600 million years to 100 million years ago there is missing geological events at Disko (Bruun et al., 2006). From cretaceous there was developed a large sedimentary basin, the Nuussuaq basin. Over vast areas there were big low lying riverbeds and deltas. These areas deposited sand, peat and in the deep oceans mud (Bruun et al., 2006). This can be seen today, on the east coast of Disko, as sandy rocks, with black bands going through them. The black bands are the peat, which over time has been changed to coal (Bruun et al., 2006).

Coal is a natural resource that has been mined in Greenland (Lyck & Taagholt, 1986). In 1924-1972 600.000 t was mined on Disko, but mining was cancelled in the 1970's when the low oil prices made it unprofitable to mine the coal (Lyck & Taagholt, 1986). In this region at the Nuussuaq peninsula an area of 325 km^2 is estimated to contain about 10^8 t of coal (Lyck & Taagholt, 1986).

In late cretaceous and paleogenic era the Nuussuaq basin was influenced by immense tectonic forces. The basin was torn up in big segments, where layered series came to tilt as much as 10 degrees (Bruun et al., 2006). In the Nuussuaq basin large marine basins emerged that has deposited fine black mud rich in organic matter (Bruun et al., 2006). The conditions in the basin were anaerobic, which caused bacteria to transform sulphate to sulphite and bind iron (Fe^{2+}) and form pyrite ($FeSO_4$). These pyrite rich shales can be found all over Disko (Bruun et al., 2006).

The deposits rich in organic matter and pyrite can lead to a rare phenomenon called "burning mountain". Oxidation can occur when these sediments are thawed. Oxidation of pyrite is known from mine trailing where reduced iron and sulphate rapidly reacts with atmospheric oxygen and produces acid (Borggaard & Elberling, 2007). This reaction releases immense heat (1100 °C) and can cause the carbon to combust. In the middle of the 1900 a little slide burned for several decades (Bruun et al., 2006).

Volcanic activity that happened at Disko 60-62 million years ago created the volcanic lava benches, which can be seen today (Bruun et al., 2006). They were caused by intense volcanic activity, the same global event that took place in Eastern Greenland, Scotland, Norway, Iceland and at the Faroe Islands. This can be seen around Godhavn as large plateaus in the landscape (Bruun et al., 2006).

Holocene deposits

After the end of the last ice age (Weichsel) the island of Disko was dominated by melting ice, which was partly covered in sediments (Humlum & Pedersen, 2006). As the ice melted, rapid eustatic sea level rise occurred covering the flat parts of Disko with water. Slowly the isostatic land rise caused the coast to regress, exposing the previously submerged landscapes (Humlum & Pedersen, 2006, Rasch, 2006).

After the last ice age the melting glaciers released massive amounts of sediments. These sediments were washed out in the ocean (Humlum & Pedersen, 2006). Coastal processes later distributed the sand on sandy beaches on Disko. Everywhere at Flakkerhuk there are many rocks and stones. Nielsen (1969) suggests that these rocks were deposited by an ice tongue advancing from the central part of Disko.

At Flakkerhuk there are several fossil beach terraces, as can be seen in Figure 2. These terraces differ in age. The oldest terraces are the ones that were exposed first aging around 10.000 years of age, and the youngest are the last to be exposed (Elberling & Pedersen, 2006).

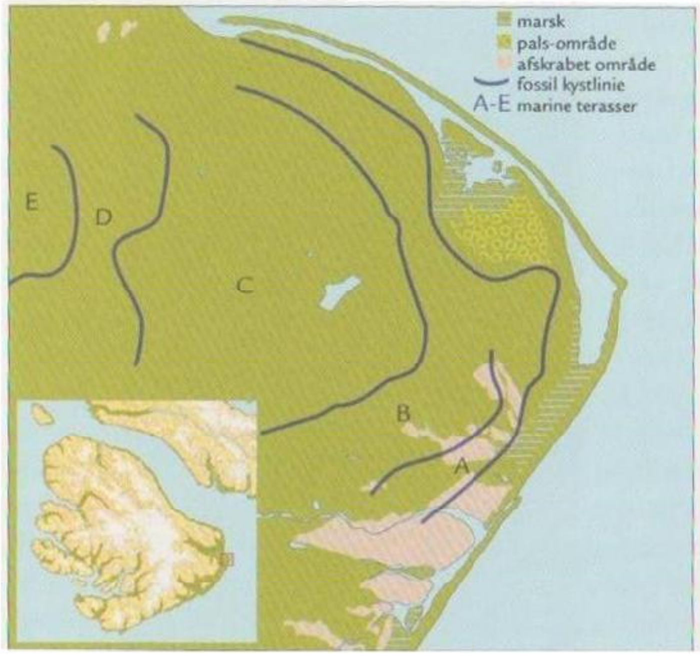

Figure 2: Map of Disko, West Greenland, showing the marine terraces in the area of Flakkerhuk.

The landscape at Disko is a mixture of glacially deposited sand, gravel and rocks, and due to the isostatic land rise much of the sand have been reorganized in the coastal zones leaving fossil terraces (Nielsen, 1969). The exposed land has later on been subjected to solifluction (Nielsen, 1969) and aeolian sediment transport (Elberling & Pedersen, 2006). An over layer of aeolian transported sands has buried old surfaces, trapping organic matter (Elberling & Pedersen, 2006).

Local climate in Godhavn and Flakkerhuk

Disko Island is located on the west coast of Greenland in Disko Bay but has also a coast line towards the Baffin Bay. The climate here is controlled both by the sea ice covering Disko Bay for several months each winter and the low pressure systems generated further south from the cold polar fronts and winds coming out from Cumberland Strait. These are sending cold and moist air to primarily the western part and large parts of the centre of Disko Island. These parts are therefore classified as maritime low arctic (Figure 3) according to the climate zones defined in Hansen et al. (2006). A more dry part of the island is found on the southeast coast which is influenced by the cold dry air coming from the sea ice in Disko Bay. Low arctic areas are defined as areas where the warmest monthly mean air temperature is between 5 and 10 °C, whereas the northern part of Disko is classified as high arctic where the maximum monthly air temperature is > 5°C (Hansen et al., 2006). Godhavn and Flakkerhuk are located 60 km apart (see Figure 1) in two different climate zones so that an introduction to the climate regime in Godhavn based on observed data is not completely representative for the study area at Flakkerhuk. However it is assumed that for an overview, the variations in climate between Godhavn and Flakkerhuk are minimized when looking at trends over the last 20 years.

Figure 3: Greenland climate zones. The subarctic zone is only found as smaller spread out areas in the inner parts of the southern Greenland fjords (Bruun et al., 2006).

Air temperature and precipitation

The main meteorological parameter controlling the climate in Godhavn and Flakkerhuk is air temperature. Figure 4 shows the annual variation in air temperature with monthly mean, minimum and maximum and is based on data from 1991-2011. For 2011 data from January to July is only included. July is in average the warmest month with a mean air temperature of 8.0 °C and the coldest month is February with a mean air temperature of -14.0 °C. There is a larger difference in air temperature between maximum and minimum values in the colder months January-March than the rest of the normal year. Figure 4 shows that the largest amount of precipitation falls in August with an average of 49 mm. The data is based on precipitation measurements at Arctic Station. The annual mean precipitation measured at Arctic Station is 255 mm, which only include rain. Based on snow fall correlated to the latitude it can be estimated that 61.5 % of the precipitation in Godhavn falls as snow. In the period 1991-2004 the average annual snow fall was ~200 mm snow–water equivalents (SWE). When knowing SWE it is possible to estimate n_t, which is a factor varying from 0 to 1 and defines the amount of freezing that reaches the surface of the ground (Hansen, 2011a). The higher the SWE - the smaller the n_t due to the insolating effect of the snow pack.

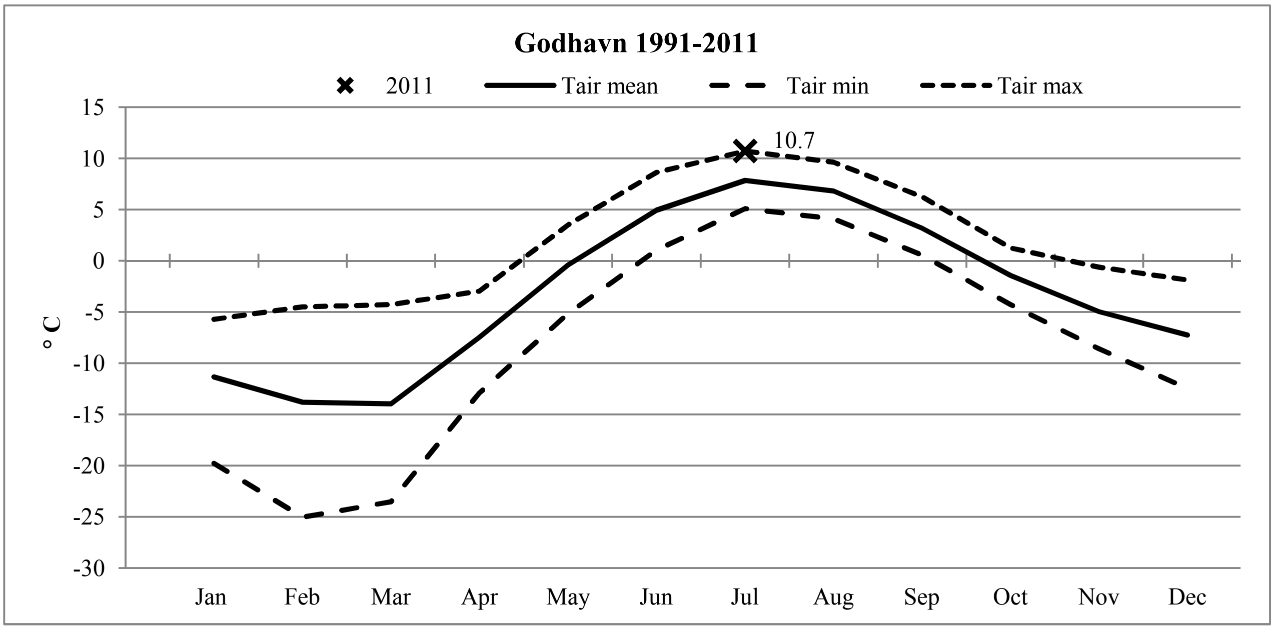

Figure 4: Annual variation in air temperature and precipitation in Godhavn (1991-2011).

An increase in temperature has been observed in the Arctic (ACIA, 2004) during the last two decades. Recorded air temperature data for July and January from Arctic Station in the years 1991-2011 shows an average increase of 0.18 °C per year for January and 0.45°C for July (Figure 5). This indicates that the warming in this area is primarily occurring in the summer months where as in winter the inter-annual variation in air temperature seems more stabile through the period. Year 2010 had an annual mean air temperature of 0.6 °C, which makes it the warmest year observed at Arctic Station since the micrometeorological measurements were initiated in 1991 and the first year where the annual mean air temperature were above 0 °C. The air temperature data for 2011 also shows an extraordinary high value for July (10.7 °C). This is the highest mean air temperature seen in July in the data series for the period 1991-2011.

The Arctic can be divided into high arctic and low arctic. A more detailed approach for dividing the Arctic area into classes in relation to the vegetation and thermal regime is the classification by Walker et al. (2005). The circumpolar area is divided into five bioclimate subzones named from A to E (see Table

1). The high arctic is represented in the subzones A, B and C whereas the low arctic is defined as subzones D and E. The subzones are defined on the basis of the summer air temperature and the dominating vegetation in the area. The definitions include July mean air temperatures (T_{July}) and a summer warmth index (SWI) which is defined as the sum of the monthly mean air temperature for the three summer months June, July and August (Walker et al., 2005). According to the classification by Walker et al. (2005) Godhavn shifted from being included in subzone D to the warmer subzone E. This is due to the fact that SWI in the last five years (2005-2010) have been 20-35, see Table 1. It is discussible whether the area is fully shifted to subzone E yet as the mean air temperature for July does not exceed 10 °C for all the years in 2005-2010 but there is a clear trend that the ecosystems in the area will change towards a different thermal regime as generally higher air temperatures are observed in all summer months.

Subzones	**Definition**	**Mean air temperature July**	**Summer warmth index (SWI)**
A	Arctic Polar Desert	1-3	< 6
B	Northern Arctic Tundra	4-5	6-9
C	Middle Arctic Tundra	6-7	9-12
D	Southern Arctic Tundra	8-9	12-20
E	Arctic Shrub Tundra	10-12	20-35

Table 1: A selected part of the specifications of the bioclimate subzones from Walker et al., 2005.

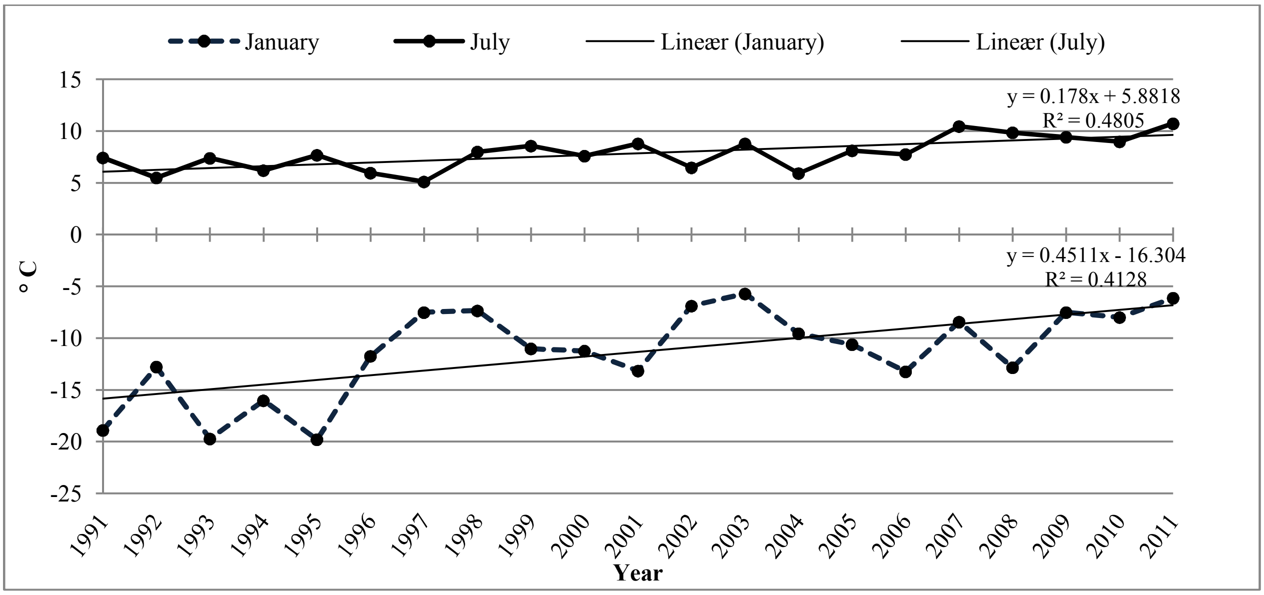

Figure 5: Mean air temperature for January and July 1991-2011.

Solar radiation and albedo

The observed incoming shortwave radiation in daily mean of the period 1991-2010 is presented in Figure 6. Due to the location of Godhavn and Flakkerhuk at the latitude 69'15° N i.e. north of the Arctic Circle there is a dark period from November 29th to January 11th where the sun does not appear above the horizon (see Figure 6). In the period 1991-2010 the surface received in average 840 W/m^2 per year. The amount of incoming shortwave radiation and outgoing radiation varies from year to year depending primarily on cloud cover and snow cover. Figure 7 shows the albedo for year 2000 - a snow rich year and 2007 where there was almost no snow. From Figure 8 it can be seen that for 2000 the snow cover melted away ultimo May whereas in 2007 the surface was snow free already medio April. This difference in snow cover duration causes a difference in the amount of energy received at the surface in the first six months (January - June) between the years. In 2000 380 W/m^2 was received whereas 505 W/m^2 was received in 2007. This is energy available for melting snow or heating the soil.

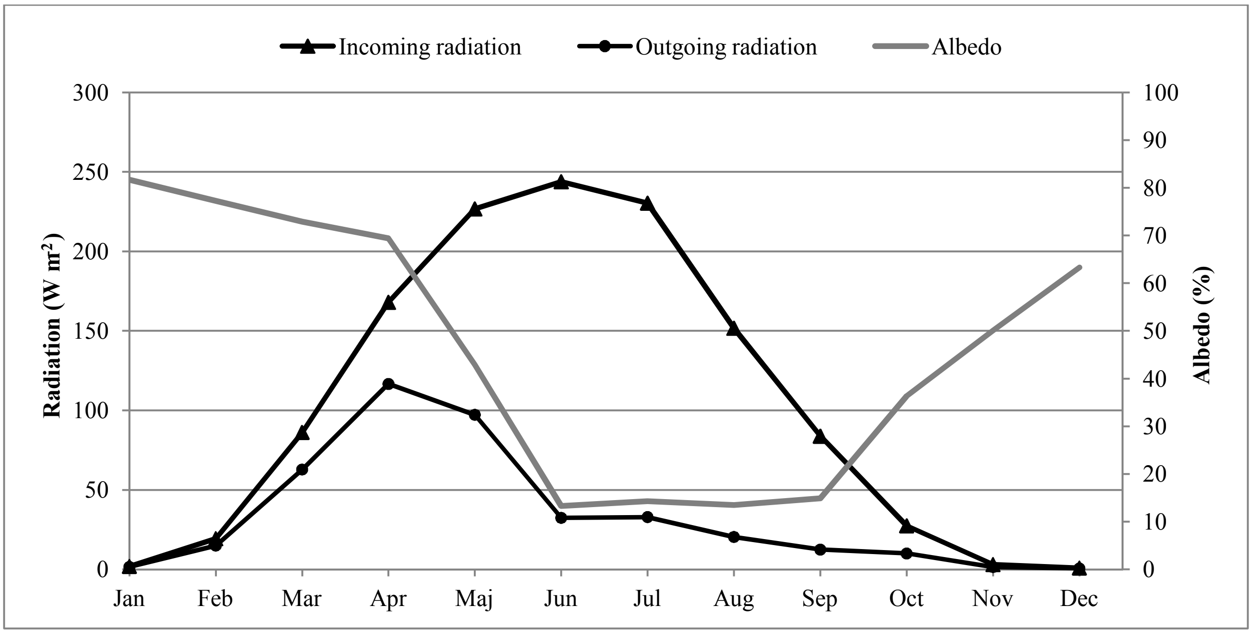

Figure 6: Monthly mean incoming and outgoing shortwave radiation at Arctic Station 1991-2010 and calculated albedo.

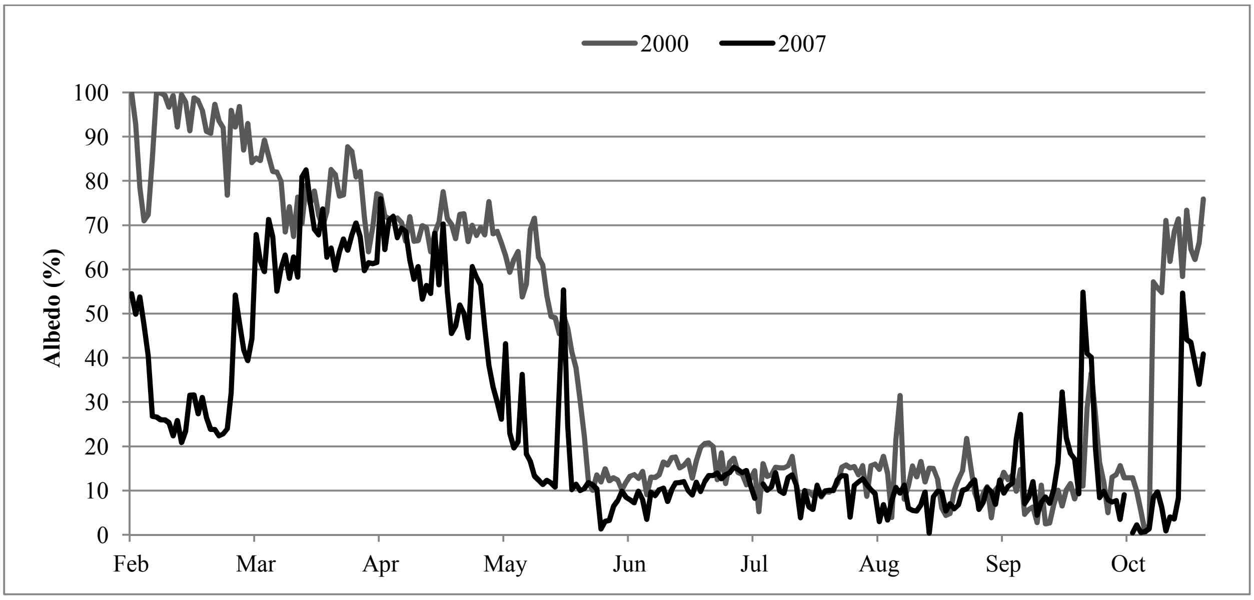

Figure 7: Temporal variation in albedo for year 2000 (snow rich) and 2007 (poor in snow).

Snow cover

The second parameter in the area controlling the climate is the cryosphere i.e. the snow cover on land and the sea ice in Disko Bay. The increase in air temperature during the last 20 years seen in Figure 5 has affected the snow cover (Figure 8). A parameter often used to determine thawing of snow is thawing degree days (TDD) which is based on daily mean air temperatures. TDD is defined as the sum of positive degrees of the mean daily air temperature over a period (Hansen, 2011a), i.e. if the daily mean air temperature is 5 °C it equals 5 TDD. The sum of TDD over a period is applicable for modelling purposes as it gives a value for the energy available for melting snow or thawing ice in the ground which can be used as a direct input parameter e.g. in a model for active layer thickness.

At Arctic Station snow cover observations have been made in the period 1991-2010 (Figure 8). The data shows a decrease in the annual duration of the snow cover over the 19-years period. The trend through the period is that the date where the surface is snow free is occurring earlier in the year and that the snow depth through the snow covered period has decreased. Low snow depths have especially been observed in 2006-2010. In these years there have not been measured any snow depths above 50 cm. The snow cover has great influence on the thermal properties of the soil as is can act as an insulating layer between the soil and the atmosphere. The snow cover can prevent frost from penetrating into the ground and thereby also for heat transport into the ground. This makes snow cover and its duration an important factor when looking at thawing rate of permafrost and active layer thickness variations on an annual basis as it defines the amount of energy available for melting ice in the ground.

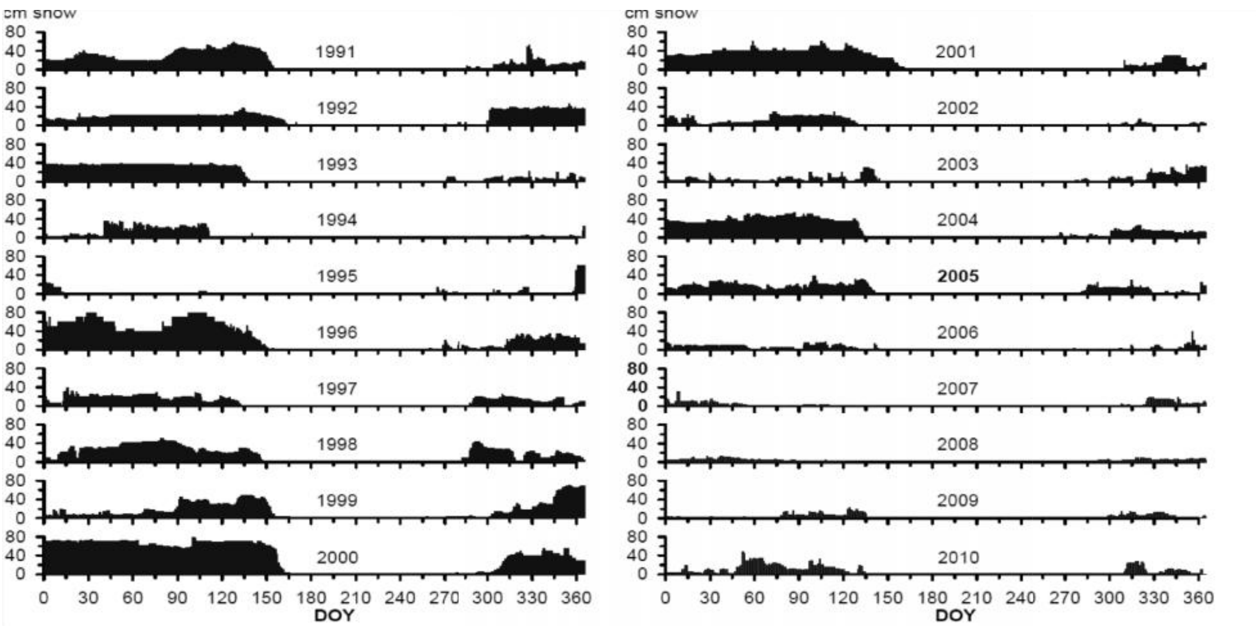

Figure 8: Daily snow depth observations in year 1991-2010 at Arctic Station (Hansen, 2011b).

Vegetation

Greenland is dominated by glaciers that cover 84 % of the landmass. The non-glacial area is evenly divided between high and low arctic zones (Walker et al., 2005). Disko is located in the transition zone between the high and low arctic. This means that the southern part of the island is the northern frontier of the low arctic species, whereas the northern part is the southern frontier of the high arctic species. These conditions results in a high diversity of plants on Disko compared to other locations at the same latitude. Of the 513 plant species recorded in Greenland half of them are found on Disko and of the West coast species 70 % is found on the island (Bruun et al., 2006). However, the arctic conditions means short growing season and low summer temperatures, which excludes tree growth and causes a low vegetation dominated by dwarf shrubs, herbs, lichens and mosses (CAVM, 2003).

Figure 9: Example of the "erect dwarf shrub tundra" physiognomic class, subzone D (CAVM, 2003).

The climate regime of Flakkerhuk is classified as continental low arctic (Figure 3), whereas the vegetation mapping by Walker et al. (2005) depicts the physiognomic class "erect dwarf-shrub tundra", which is seen in Figure 9. Walker et al. (2005) also makes use of bioclimate subzones to differentiate in the main classes according to summer temperature and vegetation. Disko lies mainly within subzone D, but the northern part is within C, which further emphasize the transition zone of the island.

However, the classifications only accounts for the dominant growth forms of the plants. The low-land zone of old marine terraces between the coast and the foothills of the mountains consist of a range of different plant communities according to the hydrological regime and topography, see Figure 10. The most seen plant community is the dwarf-shrub heath. The soil is well-drained, but not too dry and the vegetation height is rarely above 10 cm. The species are mainly Dwarf Birch (*Betula nana*), White Arctic Bell-heather (*Cassiope tetragona*), Crowberry (*Empetrum hermaphroditum*) and with slightly higher soil moisture Northern Willow (*Salix glauca*) is seen. Often the dwarf-shrub heath will have a carpet of moss below the plants. The second most seen plant community in the area is the fen. It is poorly drained or water-saturated soils where the tallest grasses can reach a height of 30 cm. It includes species like Arctic Cotton-grass (*Eriophorum scheuchzeri)*, Arctic Marsh Willow (*Salix arctophila*), mosses and different kinds of grasses. Other and less seen plant communities to be mentioned is fell-field, where the sparse vegetation is wind and draught tolerant species like Entired-leafed Mountains Avens (*Dryas integrifolia*) and tidal meadows found closest to the beach, where the soil has a high content of water, chloride and sodium and tolerant species like Creeping Alkali-grass (*Puccinellia phryganodes*) grows (Feilberg et al., 1984).

The plant community is also an indicator for emission of CH_4. Comparisons between plant communities by Elberling et al. (2008) measure no CH_4 emissions from Cassiope heath. However, major effluxes from grasslands and effluxes increasing with plant communities determined by higher water content are seen (Figure 10). Thereby detection of plant communities can be used for estimating the potential CH_4 emissions at large inaccessible areas. The different plant communities are easily recognized with NDVI or reflectance analysis using remote sensing. This can also be used for detecting changes in emissions over time as the plant communities will change due to water content and temperatures.

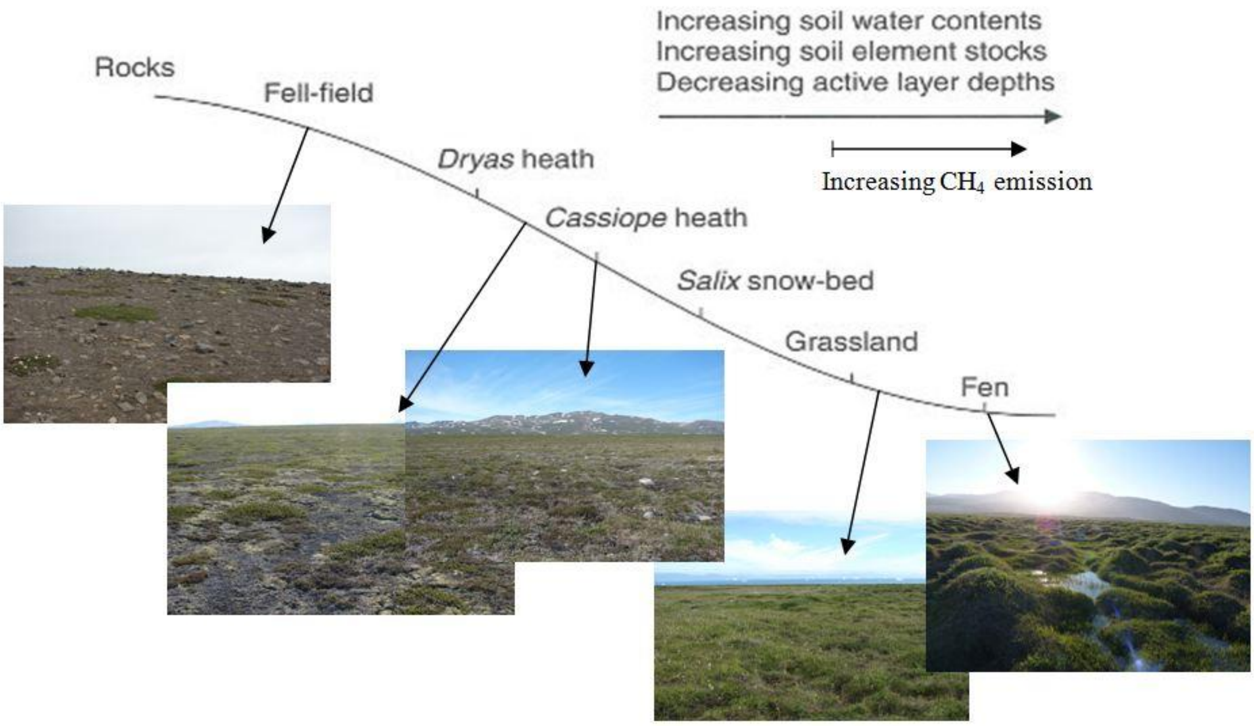

Figure 10: Plant communities and characteristics (modified from Hansen, 2011, own photos).

Methane

Methane (CH_4) processes and its effect as greenhouse gas (GHG) is subject of intense investigation in relation to climate change. Not at least because the global warming potential (GWP) of CH_4 is 25 times higher than that of CO_2 on a 100 year time scale (IPCC, 2007). Enhanced concentration of CH_4 in the atmosphere corresponds with changes in the climate, along with other important GHGs as CO_2, N_2O and water vapour. CH_4 concentrations today are around 1750 ppbv. To comparison the concentration at pre-industrial times was 700 ppb (Wuebbles et al., 2002). Currently global emissions of CH_4 to the atmosphere is around 500-600 Tg yr^{-1} (Riley et al., 2011) released through both anthropogenic and natural sources. Rumination, rice production and the use of fossil fuel are the largest anthropogenic sources (Herbst et al., 2011). Natural sources include wetlands, termites, oceans and hydrates (Petrescu et al., 2008), with wetlands being responsible for 70 % of the total natural CH_4 emission (Megonigal, 2004).

Permafrost

Temperature regime is one of the factors that through soil processes determines soil formation. The temperature regime is based on the temperature at a soil depth of 50 cm. For areas with a mean annual temperature of less than 0 °C, the temperature regime is determined as permafrost. Permafrost is used to classify the frozen soil in areas where the temperature is always ≤ 0 °C or for at least two consecutive years. In practice permafrost is a hard layer, formed by the cold climate where the ice is the most important cementing factor. The existence of permafrost is dependent on a dynamic balance between the mean annual air temperature and the geothermal gradient. When permafrost occurs within 2 meters from the surface the soil is classified as a gelisol. The active layer is the upper layer of the soil subject to freezing and thawing on an annual basis. During summer time the active layer will thaw and often turn into mud as the frozen underground prevents water from melting ice to drain away easily. Because of this, gelisols are often very unstable landscapes. Gelisols cover about 10 % of the global soils, though it is 25 % when looking only at the northern hemisphere. They are found in high-latitude regions around Arctic and Antarctica and at high mountains in lower-latitudes, i.e. the Himalaya Mountains (Borggaard & Elberling, 2007).

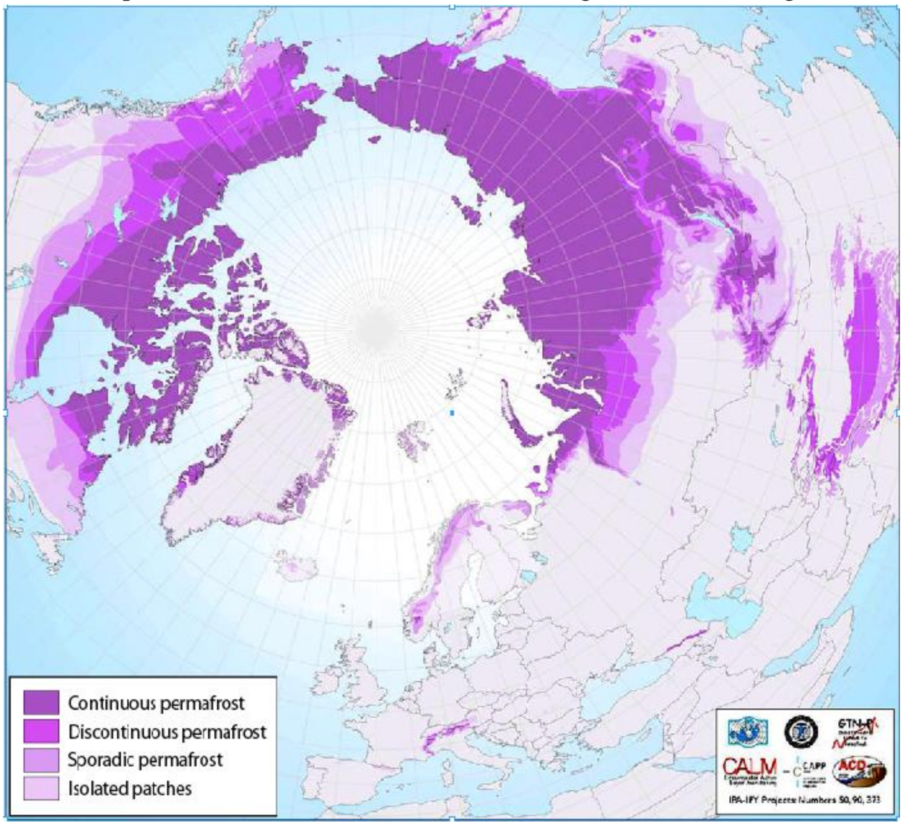

Figure 11: The four permafrost zones in the northern hemisphere (IPA, 2010)

Permafrost can be divided into four different categories as seen in Figure 11: Continuous, discontinuous, sporadic and isolated patches. The definition of the continuous permafrost is where the temperature is always ≤ 0 °C or for at least two consecutive years. The discontinuous permafrost occurs where the mean annual air temperature is only slightly below 0 °C and will form only in sheltered spots, i.e. temperature rises when protected from the wind. Usually permafrost will remain discontinuous in a climate where the mean annual soil surface temperature is between -5 and 0 °C. Sporadic permafrost is found where the permafrost covers less than 50 % of the landscape and typically occurs at mean annual temperatures between -2 to 0 °C. In the discontinuous and sporadic zones permafrost-free terrain is common. The thickness of permafrost varies from one meter to 1500 meters (IPA, 2010). All four types of permafrost are found in Greenland. At Disko the primary type found is the discontinuously. At a local scale the distribution of permafrost is further markedly influenced by factors such as snow cover, precipitation, vegetation, topography and soil type (French, 2007).

By the end of the 21st century, the IPCC AR4 GCMs project predicted an increase in the Arctic mean annual air temperature of 2.5–7 °C (relative to the period 1981–2000) under the SRES A1B emission scenario (Chapman & Walsh, 2007). Particularly, an increase in spring and summer air temperatures is expected to increase the thickness of the active layer (Osterkamp, 2005).

Methane dynamics in permafrost regions

In the Arctic region permafrost preserves labile organic matter from being decomposed, because soil processes proceeds at a minimum rate or stop completely when the soil is frozen. Global warming is observed locally at Disko resulting in temperature rising (Figure 5). The increased temperatures will melt the permafrost and lead to an increase of the decomposition rate of organic matter. Preserved carbon will be available for decomposition with additional thawing, which will result in emissions of carbon dioxide or methane. These processes are considered among the most important potential feedback from terrestrial ecosystems to the atmosphere (Dutta et al., 2006; Rodionow et al., 2006; Zimov et al., 2006).

Under anaerobic conditions, decomposition of organic matter will favour CH_4 production, see Figure 12. Therefore wetlands have high rates of CH_4 emissions. CH_4 can be released into the atmosphere by 3 different pathways: simple diffusion, ebullition and plant transport (Smith et al., 2003). However on dry sites aerobic conditions stimulates oxidation of CH_4 and emissions of CO_2 will occur instead. Both the reactions of CH_4 production and oxidation processes are seen in Figure 12. Because CO_2 and CH_4 have widely different GWP, it is of high importance which of the gasses that is being emitted.

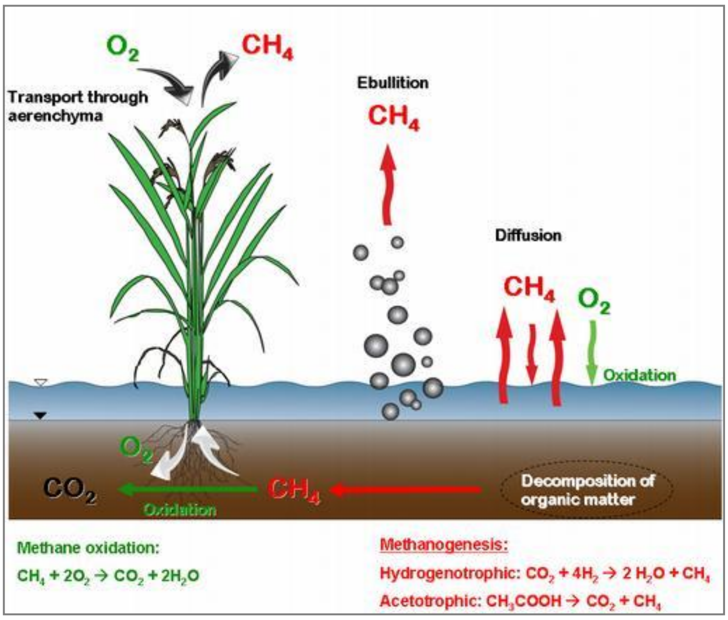

Figure 12: Transportation routes of CH_4 from the soil into the atmosphere (IBP, 2011).

Increased decomposition rates can result in changing dynamics regarding the amount of GHGs being emitted from these areas e.g. wet and dry areas. The current question today is therefore whether an increase in GHG emissions will occur or if the emission rates will be stable at the current level. An increase would amplify the climate changes already seen today, making permafrost thawing a positive feedback mechanism to increased temperatures globally. The decomposition of organic C in soils is not only responsible for a direct effect in terms of emitted CO_2 and CH_4 to the atmosphere, but can also be accompanied by a subsurface heat production which can result in a positive feedback on soil temperatures leading to further soil thawing (Khvorostyanov et al., 2008).

Because of differences in thermal conductivity between wetlands and dry sites an increase in temperature will not have the same effect on the two sites. As it takes less energy to increase the depth of the active layer in a dry site, the permafrost boundary will be lowered more than during the same temperature increase for a wet site. Yet it is foremost the water table depth and soil water content that determine whether CH_4 production or consumption is the dominant process within a given ecosystem. CH_4 consumption can occur under aerobic conditions because of oxidation. Both processes are also influenced by soil temperature, pH and vegetation. A change in these parameters will therefore affect the processes in terms of surface flux rate.

If the permafrost is carbon rich it will increase the potential production of GHGs thereby enhancing the climate changes as a positive feedback mechanism. This call for a discussion of what will have the largest effect on climate change: a deep thawing at a dry site resulting in increased emissions of CO_2 or a smaller thawing of a wet site resulting in increased CH_4 emissions.

Article overview

Considering the driving factors for methane dynamics in permafrost regions mentioned above, the following articles are conducted to investigate such dynamics on a local scale for Flakkerhuk, Disko Island.

Soil serves as storage for carbon (C) and nitrate (N). Hence, soil characteristics play a major role, as changing weather conditions can lead to large changes in the pedological ecosystem storage functions, which can either lead to emission or uptake of GHGs. It is therefore necessary to examine the soil properties at Flakkerhuk to investigate what threat this near-coastal landscape poses to global change. The first article *´Comparing soil types and underlying permafrost within a near-coastal landscape at Disko, Greenland´* is describing soil characteristics at Flakkerhuk. The scope is to determine chemical and physical dynamics by looking at texture and water content and measure C and N in the solid phase and the water phase. By looking at C/N ratios a rough indicator of growth potential for micro organisms and plants is achieved.

A driving factor for pedological dynamics, especially soil thawing is climate. Therefore the local climate in Flakkerhuk is analysed. Special attention is drawn on the spatial climate distribution which is presented in the second article *'Climate gradients at Flakkerhuk, Disko'*. The main focus is on temperature regime as a result of the energy balance of the area. Due to the energy balance, contributing parameters such as solar radiation, humidity and wind need to be related and compared to the observed temperature trends in the area. A transect sample strategy and a permanent climate station in a mountainous area can reveal potential differences between coast and inland areas and gives the ability to show the importance of the laps rate in temperature trends of landscapes.

In order to better understand the CH_4 dynamic dependencies, investigating the role of different surface types and site specific parameters such as vegetation type, soil water content and soil temperature is important. The article '*Methane fluxes measured at Flakkerhuk, Disko Island (West Greenland), at different vegetation specific sites'* presents such a study of CH_4 fluxes measured on two wet and two dry sites. The fluxes are presented and stressed in terms of emission and uptake.

With the site specific CH_4 fluxes measured at wet and dry sites an upscaling is enabled. Upscaling is interesting in terms of CH_4 net balance hence, whether the area is a sink or source to atmospheric CH_4. In the article *'Upscaling methane fluxes to a net balance for Flakkerhuk, Greenland'* the fluxes are upscaled to Flakkerhuk aiming to perform an estimation of CH_4 net balance for midsummer in Flakkerhuk.

In order to compute landscape specific CH_4 production under changing climatic conditions a permafrost model is required. Permafrost and summer thawing strongly depends on the temperature regime in the air and soil, but also on thermal properties of the soil. By implications thermal properties are closely related to the physical properties of soils and its water content. This relation is described and analysed as well as the model is set up in the article '*Model the effect of active layer thickness in arctic soils under changing moisture conditions'*

Finally to evaluate the future impact of active layer dynamics the article '*Climate change feedbacks of the future greenhouse gas budget'* is focusing on future climate scenarios and how increased temperatures will affect CH_4 fluxes. The article applies results from the other articles and the future climate scenarios used are 0.25 °C and 5 °C. Coupled with CH_4 oxidizing potential the article gives an estimate of potential GHG emissions under increasing temperatures for wet and dry soils within Flakkerhuk.

An overview of the order of articles is listed below.

1. **Comparing soil types and underlying permafrost within a near-coastal landscape at Disko, Greenland. Page 21-30.**

2. **Climate gradients at Flakkerhuk, Disko. Page 33-41.**

3. **Methane fluxes measured at Flakkerhuk, Disko Island (West Greenland), at different vegetation specific sites. Page 43-55.**

4. **Upscaling methane fluxes to a net balance for Flakkerhuk, Greenland. Page 57-61.**

5. **Model the effect of active layer thickness in arctic soils under changing moisture conditions. Page 63-77.**

6. **Climate change feedbacks of the future greenhouse gas budget. Page 79-86.**

Summary

The study area of CH_4 dynamics in a permafrost landscape is located in Flakkerhuk, Disko Island. The area is characterized as a near-coastal landscape, which has undergone isostatic land rise and formations of fossil beach terraces. Flakkerhuk is influenced by an arctic climate with July being the warmest month of 8.0 °C and February the coldest with -14.0 °C. Over the years 1991-2010 an average increase in mean air temperature of 0.25 °C per year has been recorded and largest amount of precipitation falls in August with 49 mm. The yearly snow cover is decreasing both in terms of duration and thickness. The duration of snow cover and the isolating properties are important for the thawing rate of permafrost and the active layer thickness as it determines the energy available for thawing the ice in the soil. In terms of vegetation Flakkerhuk is characterized as continental low arctic due to the climate and short growing seasons. There is a high diversity in plant species which occur in a mosaic pattern. The dominating vegetation types are low dwarf-shrub heath and fen.

The studies in this report clarify the current environment in Flakkerhuk and the interaction between biogeophysical parameters and importance on the soil-air dynamics. A spatial climate distribution with a temperature gradient that increases exponentially when going inland was found. The distance to the coastline is the driving factor more than the lapse rate. Regarding soil properties wet soils are found to have higher concentrations of carbon (C) and nitrate (N) than dry soils, which were also expected. The C contents are highest in the top of the profile and rather low further down. However, in general the concentrations of C and N are low. The difference in soil properties and vegetation cover on wet and dry soils respectively created expectations of different CH_4 fluxes from these soil types. Emissions from wet soils of on average 3.74 mg/m^2/day and uptake on dry sites of on average 1.31 mg/m^2/day are successfully recorded. Correlations between CH_4 fluxes and soil water content and soil temperature were expected to be found yet this is not the case. The lack of correlation might be due to the short period of measurement hence, the results show a snap shot of the arctic midsummer conditions. When upscaling the flux values to Flakkerhuk, where the spatial distribution of wet and dry sites is estimated to be 10 and 85 % respectively, the fluxes are 9.86 g/day vs. 30.3 g/day respectively. This gives a net balance for CH_4 of app. 20 g/day. Flakkerhuk is thereby a sink of atmospheric CH_4. Moreover, applying a temperature and moisture dependent permafrost model it is shown that the thermal properties of the soil type have a large affect on its melting potential during the arctic summer. In wet soils the thawing potential is reduced due to higher energy required in order to heat the soil column. Thus the computed values can display the dry sandy soil with an offset of 10 cm compared to the observed active layer depth. Future temperature scenarios are set to an increase of 0.25 °C and 5 °C for the arctic region however, combined with the permafrost modeling it show low CH_4 potential of the area. That is, despite the increased temperature which will extend the snow free period and enhance the permafrost thawing, a potential increase in CH_4 production is rather low due to the low C content. Based on this, Flakkerhuk is presumably continuing to be a sink of CH_4 in the future.

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Comparing soil types and underlying permafrost within a near-coastal landscape at Disko, Greenland.

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Abstract. Fieldwork was carried out at Disko Island, Greenland in July 2011, with the goal to obtain samples of deep soils (0,7m +) that are permanently frozen, and of shallower soils in the active layer. Samples were collected at both wet and dry soils**.** The aim is to determine several chemical and physical factors of the soils in the laboratory, to see how the difference of thawing wet and dry soils may affect future greenhouse gas budgets. Factors include: texture, water content, percentage of carbon and nitrogen. Permafrost samples were collected using a handheld motorized diamond drill **(**STIHL BT 121**)**. Depth-specific active layer samples were collected using volume-specific rings. The wet soils had carbon and nitrogen contents that was approximately twice as high as the dry soils. Even so the soils only contained relatively small amounts of carbon and nitrogen, so a thawing of this area would only lead to a small amount of GHG-emission. The study was very local, so therefore a spatial geographical upscaling of these results should only be applied to Flakkerhuk.

Keywords: Disko, Greenland, soil characteristics, permafrost, nitrogen, wet soils.

Introduction

To understand the processes responsible for creating a landscape it is necessary to take a look at the long term geological perspective. All landscapes in the arctic zone are affected by the presence or past presence of large glaciers or inland ice. The less than 12000 year old (Holocene) landscape at Flakkerhuk is a result of glacial- and coastal activities since the last ice age (Humlum, 2006). When the last ice sheet retreaded, the sea level rises first followed by a slower responding land mass rebound as a response to the reduced pressure. Current and past events as sandstorms or floodings has occurred and likewise lead to over layering of former top layers. This can lead to buried layers in the present landscape (Krüger, 2000). Burials of this kind can be so thick that the result is a formation of a new surface, i.e. a new top layer is developing.

Soils consist of a mixture of particles of different mineralogy, size and shape. Because the size of the particles has a significant effect on the soil behavior, the grain size and grain size distribution are used to classify soils (Nielsen et al., 1990, Jensen et al., 2001). The texture depends on whether the soil has been transported by glaciers, glacier floodings, wind or water. If transported and deposited by wind (aolian) or water (alluvial) the soil texture is fine and well sorted and contain sediments like silt and sand (Nichols, 2009). If the soil is a deposit of glaciers or glacier-floodings it is an unsorted mass of sediment (till) containing clay, silt, sand, gravel and even boulders (Krüger, 2000).

The soil forming factors, pedogenesis, is the dominating processes controlled by time, relief, hydrology, parent rock, climate, fauna and flora. These last three have a profound influence on soils. Climate affects the vertical movement of water and minerals which lead to the formation of the horizons in soils. Faunal and fungal activity breaks down organic materials mixing the soil and plant litter and determines the nature of humus. Soil temperature is one of the crucial factors for these processes to run (Bardgett, 2005). In permafrost-affected soil systems, cryoturbation can have a marked impact on the landscape formation. Cryoturbation refers to the mixing of materials from various horizons of the soil right down to the bedrock due to freezing and thawing alternately.

Both chemical, physical and biological mineralization of organic matter is of critical importance for ecosystem function since it directly determines the availability of nutrients to plants. Mineralization transforms organic matter to inorganic nutrients which can readily be consumed by plants and microorganisms (Brady & Weil, 2006). The opposite reaction is immobilization, where organic matter is absorbed by micro organisms, making the nutrients inaccessible to plants (Brady & Weil, 2006). Mineralization and immobilization can occur at the same time. The ratio of carbon to nitrogen (C/N) is one among other factors influencing whether the process results in net mineralization or net immobilization (Jensen & Jensen, 2001). The net N mineralization occurs when microbes are predominantly C-limited. In the arctic areas, i.e the cool grasslands and the Russian taiga the C/N relationship is earlier found to be around 13,6 and in a tropic forest it is found to be around 24,9 (Aitkenhead & McDowell, 2000).
At anaerobic conditions decomposition reduces soil organic carbon (SOC), producing methane in the wet soils. Decomposition of organic material at aerobic conditions favours oxidizing SOC producing carbon dioxide in the dry soils.

The aim of this paper is to investigate how the landscape was developed and to find out the conditions and properties of the soil and to find any differences of carbon stocks in the wet & dry areas as well as in the active layers & permafrost. This knowledge is important for predicting future decomposition and greenhouse gas budgets upon permafrost thawing.

Materials and methods
Five sites were chosen in the wetlands, see example of a site at picture 1. At each site two active layer samples were taken with a pipe (Ø6.8 cm) pressed into the soil before digging a hole for drilling into the permafrost. A vacuum was applied with tape before the pipes were pulled up, to ensure the entire content to be collected. The sample was afterwards pushed out of the tube and cut into depth-specific samples of 1 and 5 cm length in the top and bottom respectively. Two of the sites could not be used because there were too many stones.

Picture 1, Wet site no. 4. The dominating vegetation on this picture is carex sp. Picture taken by the group

Picture 2, Dry site no. 3. Collecting samples from the active layer prior to drilling. The dominating vegetation on the dry sites was willow (Salix glauca). Picture taken by the group.

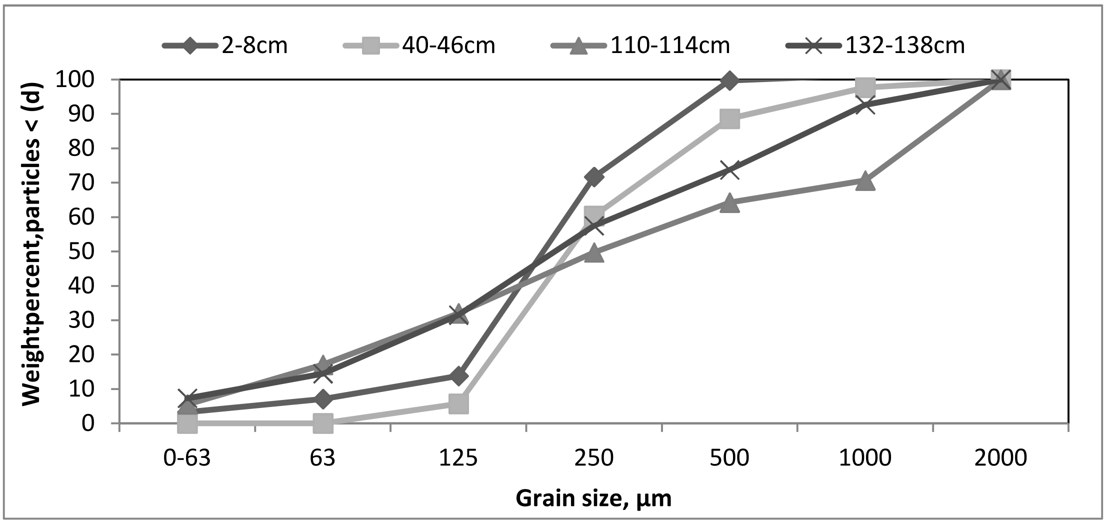
Three sites were chosen in the dry areas, see picture 2. Active layer samples were collected with volume-specific rings at each site. The volume-specific samples were taken from each horizon in both sides of the profile. The holes were dug deeper than the water table down to the permafrost. To prevent sediments from caving in from the sites and covering the drill hole, a wooden box was made and placed with the drilling hole in the center at all drilling sites.

Afterwards all active layer samples from the wet and dry sites were kept cold in a refrigerator.
Permafrost drilling was done using a diamond drill (STIHL BT 121 (4YL38), Ø 5.5 cm). The maximum depth reached was 1.95 m limited by the amount of larger stones. The permafrost cores were forced out of the drill by hammering on the core catcher, bagged and labeled as quickly as possible. The samples were kept frozen afterwards.

In total 28 active layer samples and 23 permafrost samples at different depths were selected to represent the 3 wet and 3 dry sites in the further analysis. From each of the 51 samples 20 g were collected, weighted, dried and weighted again for analysis of water content and bulk density.

In the following all samples were sieved in a 2000 μm sieve. The part of the sample with a fraction below 2000 μm was crushed and analyzed for the total carbon and nitrogen content in the LECo procedure. LECo is a three phase analysis cycle containing purging, combustion and analysis. The sample is purged for atmospheric gases, burned in furnace at 950 °C, flushed with oxygen very rapidly and combusted completely. The product is placed in an afterburner at 850 °C. In the third phase it is analyzed with a CO_2 detector and after the gases have equilibrated carbon is measured as carbon dioxide by the detector. A thermal conductivity cell is used to determine the nitrogen content (Tabatabai & Bremner, 1970).

Figure 1, Grain size distribution of the dry site



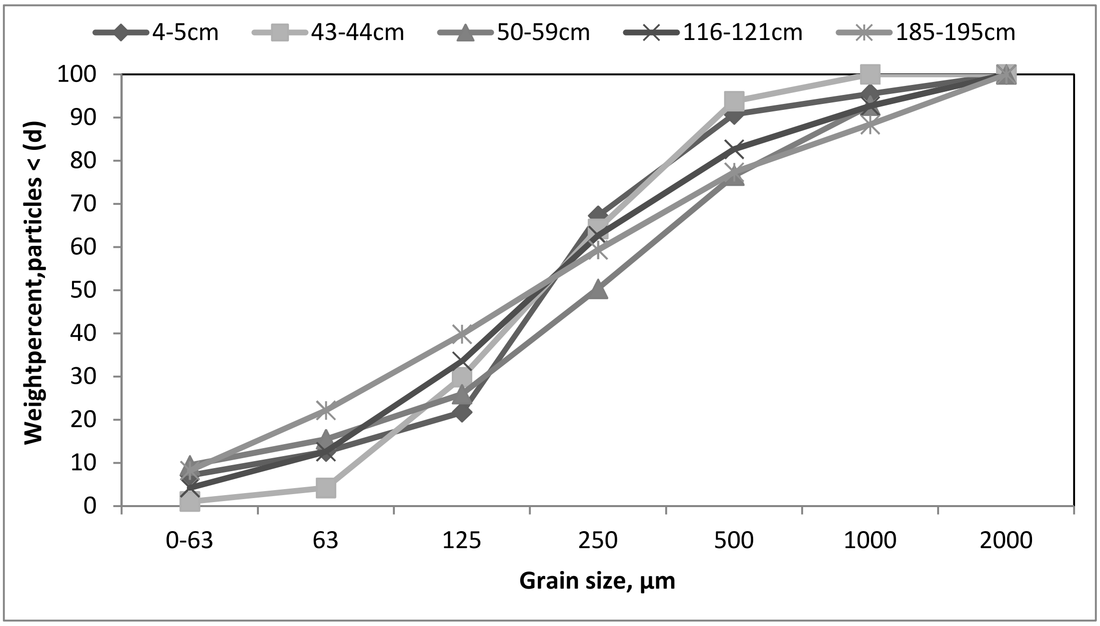


Figure 2, The grain size distribution of the wet site

Calculation of the carbon and nitrogen pools is done as follows: The densities (ρ) are multiplied with percentage of mass (m) C and N respectively, and summed for the depth of the profiles (dx). The total mass of C and N is calculated for both the wet and dry soil:

Carbon or nitrogen pool =

$$\sum(dx * \rho_{soil} \square m_{C\ or\ N})$$

All permafrost samples were analyzed for dissolved organic carbon (DOC), total N, ammonium (NH_4^+) and nitrate (NO_3^-). The samples were shaken for 15 min with deionized water and centrifuged before the water was collected and frozen. The frozen samples were analyzed in the FIA Star 5000 System (Flow Injection Analyzer). The flow injection analyzer is used for automatic wet chemical analysis of nutrients and other parameters in water, soil, plants, food products etc (Foss.dk).
Six samples from each of the sites (depths) were selected for grain size analysis. The samples were sieved for 20 minutes. The six sieves used were 2000 µm, 1000 µm, 500 µm, 250 µm, 125 µm and 63 µm. The mean grain size and standard deviation was calculated using the moment method (Nielsen & Nielsen, 1990).

Results

Grain size

An accumulated curve of the grain size distribution has a specific steepness. This steepness is a result of how well sorted the material in the soil is. A steep curve means the material is well sorted, i.e. one main grain size present in the soil. A well sorted soil is what we see in the top layers of the dry site (Figure 1). The peak i.e. the main grain size present, is in the 0.25 mm fraction and therefore the curve is steeper at this grain size. The maximum potential for sediment transported by aeolian processes is around 0.5 mm (Nichols, 2009). The texture in the top has a mean grain size of 0.46 mm (USDA = medium sand) which is increasing with depth to 0.65 mm (USDA = coarse sand). At the wet sites (Figure 2) the mean grain size is lower than at the dry site. In the top the mean size is 0.41 mm (USDA = medium sand) and increases to 0.5 mm with depth from 50-100 cm. Compared to the dry site the wet site is more evenly distributed down through the soil or in other words sorted in the same way. In the dry replica profiles the active layer consists of an A_1 and a B_1 horizon overlaying an A_2, B_2 and C horizon. The changes in grain size distribution down through the dry site could indicate that there has been an over layering of aeolian sediments. This implies that A_1 and B_1 were deposited by aeolian processes (Picture 3).

Picture 3, Dry profile. Horizons: A_1, B_1, A_2, B_2 and C.

Water content

The water/ice content of the samples is relevant when dealing with processes depending on aerobic or anaerobic conditions. The water/ice content of the wet profile (Figure 3) indicates approximately 60 % in the top 10 cm. The content is decreasing with depth to 35 % in 20 cm and only 20 % in most of the frozen part of the soil. The high water content above the permafrost is due to that the frozen soil prevents water from percolating further down into the soil (Brady & Weil, 2008). The dry site contains

20 % water in the top layer and decreases to 5 % in the depth of 15-25 cm but increases slowly with depth and contains 10-20 % in the permafrost. The two profiles contain approximately the same amount of water (20 %) in the deepest part of the profiles. The three times lower water content at the surface of the dry site could be due to a run off effect towards the sea in these areas.

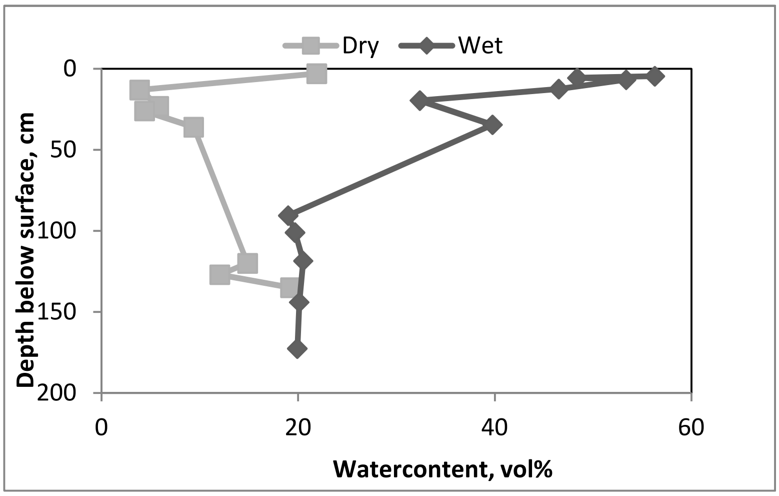


Figure 3: Water content in a wet and dry profile

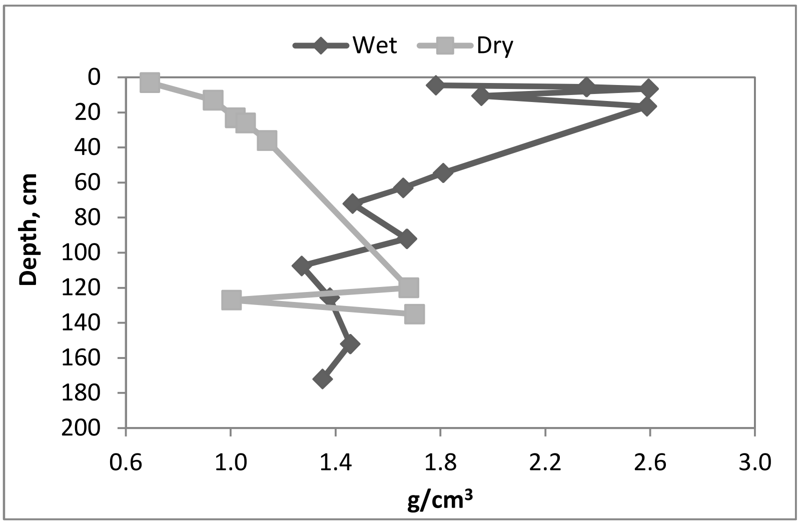


Figure 4, Density in a wet and dry profile

Density

The density is relevant when looking at soil processes as it determines the space for air and water. The density for mineral soil and organic matter is 2.7 g/cm^3 and 1.3 g/cm^3 respectively. Normally the amount of organic matter decreases with depth because of decomposition and lack of additional input and therefore the soil density increases with depth (Jensen & Jensen, 2001). In figure 4 the densities of the wet and dry profile is shown.

The density is very low in the top layers of the dry site (0.97 g/cm^3). This is due to either a high content of organic matter or a very loosely packed sand layer. The density is getting higher with depth due to compaction. In the permafrost the density is 1.4-1.7 g/cm^3, probably due to compaction both from the above layers and compaction due to frost. The density of the wet site is opposite of expected, i.e. the top layers have a higher density than the deeper layers.

Carbon and nitrogen

The dissolved organic carbon (DOC) defines the carbon content in the soil solution.

These molecular species varies between different soil solutions and are often partly associated with cations, e.g. Ca, Mg, Al or Cu (Elberling & Borggaard, 2007). The dry soil contains very little DOC (Figure 5) with a maximum in the depth 110-140 cm whereas the wet soil contains higher amounts of DOC down through the whole profile. The content rises through the wet profile from 2 ppm in the 55cm depth to the maximum content of 23 ppm in the depth of 150 cm.

At 195 cm the content is about 10 ppm. The amount of DOC in the wet soil is about 50 % higher compared to the content in same depths

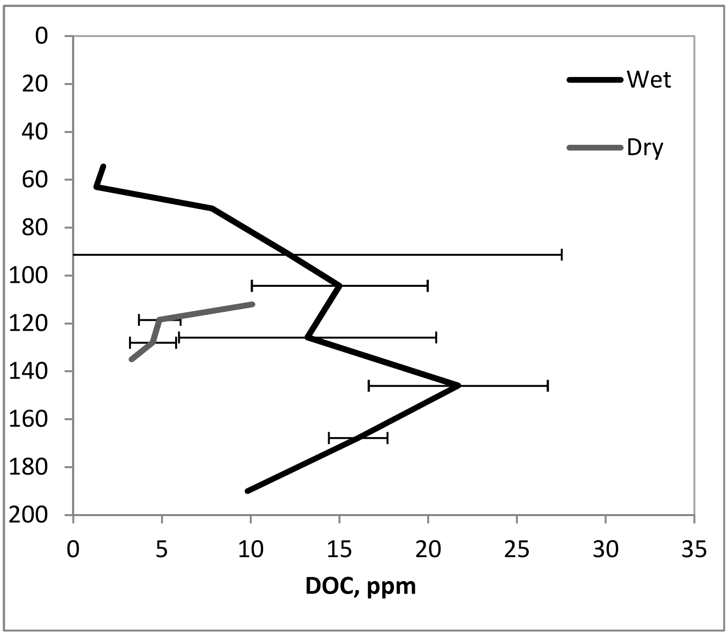


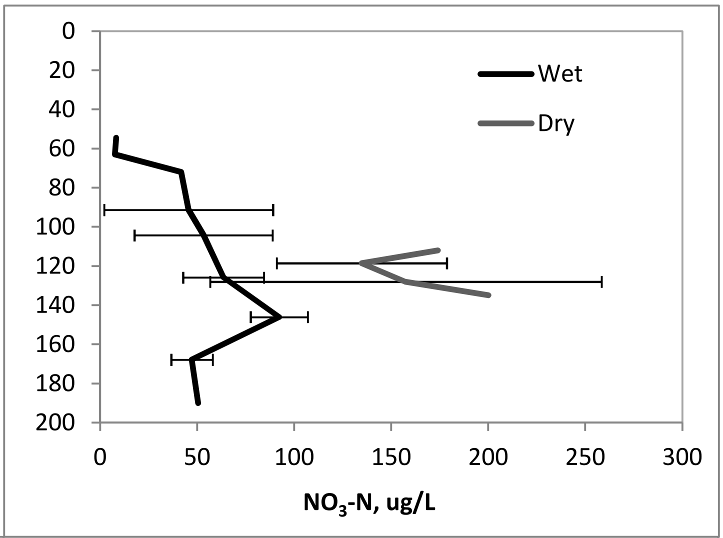
Figure 5, Depth specific dissolved organic carbon

of the dry soil. Nitrogen is converted to plant available nitrogen compounds after it is fixated as atmospheric nitrogen (N) and incorporated into organic compounds. This is a very important microbial process in relation to plant growth. The dissolved inorganic N (nitrate (NO_3^-) & ammonium (NH_4^+)) is now ready for plant uptake or the process of denitrification. The difference in DOC for the wet and dry profile can also be found when comparing total N (figure 6). The content of nitrogen is generally higher in the wet profile with concentrations ranging from 100 µg/L to 2200 µg/L. The maximum content is found in the depth of 105 cm and 115 cm for the wet and dry profile respectively.

The total nitrogen content is a combination of the content of primarily NO_3^- & NH_4^+. Nitrogen sources can also be amino acids, R-C-NH_2 or bonded to carbon (Brady & Weil, 2008). The content of NH_4^+ is for both the dry and the wet site the dominating component (Figure 7 & 8). NH_4^+ is represented down through the profile and accumulates in the same depth as total N for both the wet and dry site (at 105 and 115cm). The content of NO_3^- and NH_4^+ in the dry profile is only represented in a shallow depth interval, as only a few permafrost samples were available.

The concentration of the NO_3^- is very small with a maximum of 200 µg/L compared to NH_4^+ with a maximum of 3300 µg/L. The

Figure 7, Depth specific content of nitrate (NO_3^-)



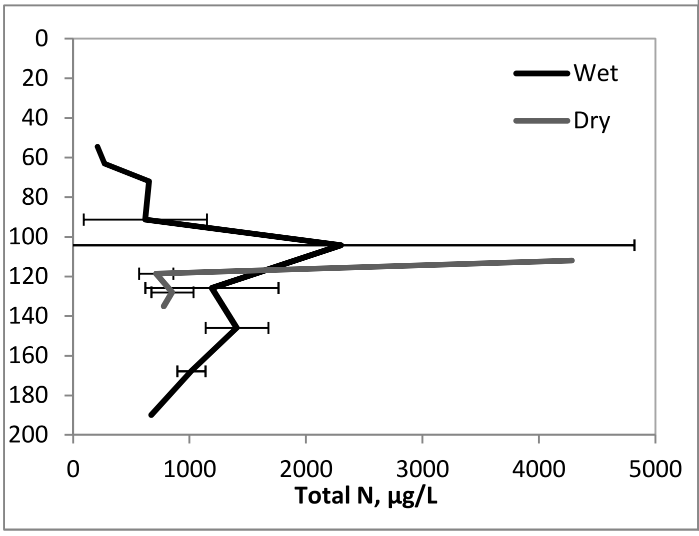


Figure 6, Depth specific total amount of nitrogen

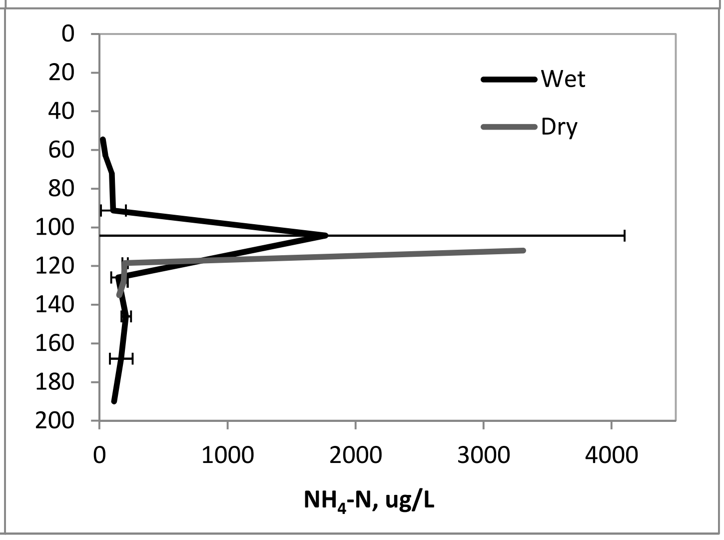


Figure 8, Depth specific content of ammonium (NH_4^+)

NH_4^+ can bond to negatively charged colloids, making NH_4^+ less leachable than negatively charged NO_3^-.

The higher amount of both NO_3^- and NH_4^+ in the depth intervals 115-145 cm in the dry compared to the wet profile is the opposite of what we expected due to the higher decomposition in the dry soils.

The higher concentration of both NO_3^- and NH_4^+ in the dry sites at 115-145 cm, can be the result of accumulation.

Carbon and nitrogen ratio

Microbes need carbon for building essential organic compounds, and to obtain energy for life processes. They also need nitrogen to make amino acids, enzymes and DNA and other cellular components (Brady & Weil, 2008).

The carbon content in the dry profile (Figure 9) is in the top layer 2 %, and decreases to 0.1 % at the depth of 25 cm and remains constant throughout the profile. The wet profile has a carbon content of nearly 4 % in the top layer and decreases quickly to about 0.3 %. This content varies down through the profile between 0.1 and 0.3 % carbon. The content of nitrogen has more or less the same distribution as carbon (Figure 10), but with maximum concentrations of less than 0.3 %. The dry soil has a maximum of 0.12 % in the top soil which decreases with depth to about 0.03 at 20 cm and throughout the profile. The wet site contains 0.28 % N in the top soil which already at a depth of only 5-7cm has decreased to 0.1 %. This content varies between 0.1 % and 0.3 % with depth.

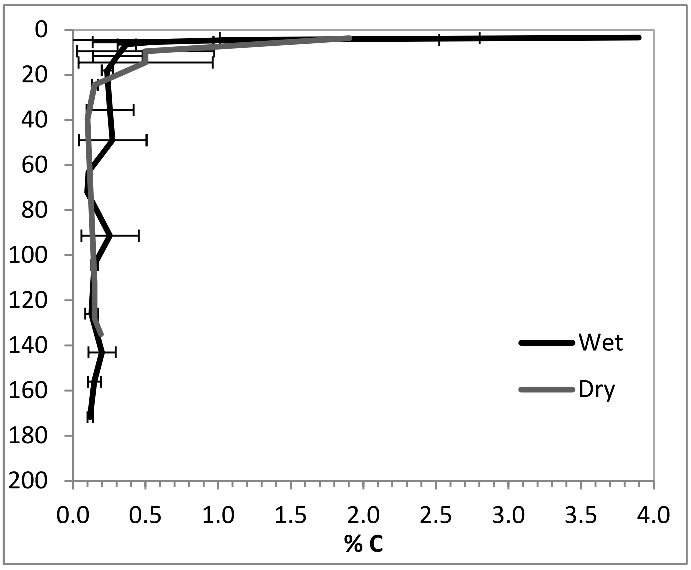
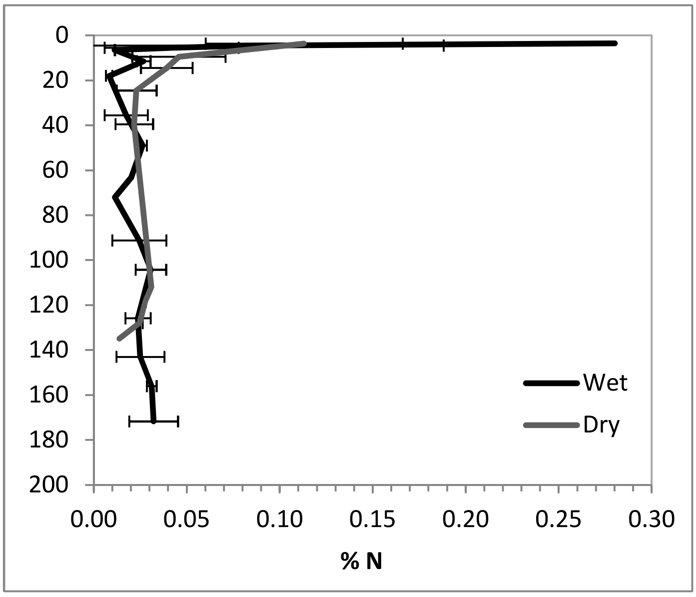


Figure 9, a full profile of the total carbon content for the wet and dry site

Figure 10, a full profile of the total nitrogen content for the wet and dry site



C/N ratio variations in the wet and dry soil

A C/N ratio of approximately 8 is optimal for fast decomposition. Therefore a high C/N ratio can result in slower decomposition rates. At high C/N ratios nitrogen is unavailable and it will be harder for plants and microbes to obtain

nitrogen (Brady & Weil, 2008).

At Flakkerhuk, the C/N ratios are high in the top (figure 11), both for the wet and dry sites. In the first 20 cm, the dry C/N is above 10. Below this the ratio is consistent at about 4 - 6. In the lowest measured point, at 138 cm the value is about 14. In the wet sites, the values are higher than in the dry.

Differing from 14 to 35 within the first 20 cm. Then decreasing values till 60 cm to 5 and rising to 10 at 90 cm depth and dropping down to 5 just below 100 cm. Rising to a value of 8 from

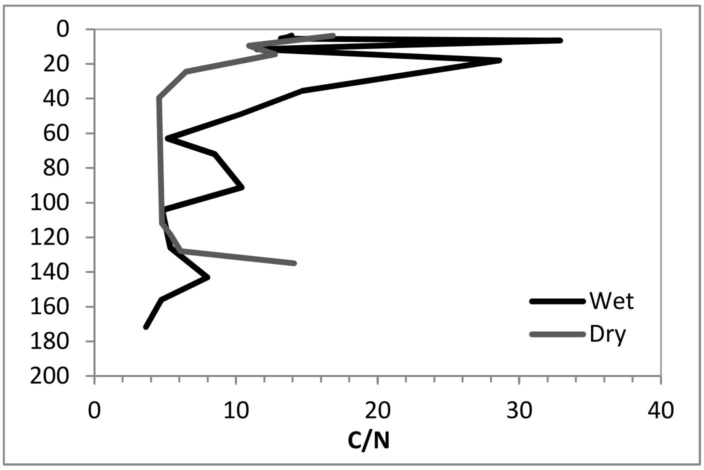


Figure 11, Depth specific C/N ratio

100 cm to 140 cm depth and then decreasing again. Though the ratios are fluctuating, the values are very low. Thus a little change in the %N can result in a large change in the C/N ratio. The results of the four pools are as follows in table 1.

	N Pool [g/m^2]	σ	C pool [g/m^2]	σ
Dry [0-150 cm]	5.48	0.45	47.22	0.98
Wet [0-200 cm]	11.15	0.91	105.04	1.23

Table 1, Total amount of N and C in the profiles.

The wet soils contain approximately twice the amount of both C and N as the dry soils (Table 1). There is about ten times as much carbon than nitrogen in the soil. The standard deviation is in generally very small which means that differences within the replicates with respect to C and N are small.

Discussion

There were a lot of rocks, which disturbed the drilling. These rocks, was present at all depths in the soil, some places more concentrated than others. This is likely to be residuals from previous geological environments.

The texture of the wet soils, are less sorted than the dry soils and also has a higher density. This coincides with the fact, that unsorted soils have interlocking grains. The interlocking grains can fill the cavities in the soil thereby raising density. The top layers of the different soils, has very similar textures, with the bulk of particles in the fraction of aeolian transported material. This illustrates that the top layer (A_1 & B_1) is created simultaneously by the same process/event.

The water content of the samples in the wet area are low compared to what would be expected in a water saturated soil. The active layer samples may have been subject to drainage and evaporation during sampling at further handling (transport, lab work ect.). As well we expected the same content of water though out the wet active layer, as it was saturated, and not decreasing all ready in the depth of about 15-20cm.

At the dry site the water content in the top sample are higher than below. This may be due to the high organic matter content which can hold more water per cm^3 than sand (Borggaard & Elberling, 2006).

The frozen samples may have lost substantial amounts of water during treatment, but less than the active layer samples, since the water is frozen and therefore less of the water will evaporate before weighing.

The density of the wet site is opposite of expected, i.e. the top layers have a higher density than the deeper layers. Density should be relatively low in the wet topsoil and higher in the dry topsoil, because the wet soils have larger grain sizes. The water content is only measured to 60%, but due to the water table in 0cm at the wet site, this is not the case.

The higher concentrations of DOC in the wet soil can be explained by an increased decomposition of DOC in the dry soil due to higher temperatures and more oxygen available. Large standard deviations do occur, but since the standard deviation for the wet and dry profile does not overlap compared in the same depth, the data is still trustworthy.

Figure 7 and 9 show the total amount of nitrogen, and content of ammonium with depth. What is of particular interest is the peak they share at approximately 110 cm depth. At this depth the active layer has reached its maximum depth, leading to an accumulation of dissolved nutrients that percolates down through the soil but is stopped by the frozen layer. Also notice that the sum of the ammonium and the nitrate is far less than the total amount of nitrogen found in the soils. This means that the soils contain a lot of nitrogen bound in other forms, than ammonium or nitrate.

In figure 10 and 11 we clearly see that the bulk of total carbon and nitrogen is being compiled in the top of the profiles. At the top layers, the basis for life is met, because sunlight provides energy for photosynthesis.

The carbon and nitrogen pools shows that there is the double amount of carbon and nitrogen in the wet soil than in the dry soil.

The vertical distributions of C and N in the profiles show a clear trend. And even though the wet soils has a higher total mass of C and N, than the dry soils, the trend is the same: the content of C and N is compiled in the top of the profile (approx. 25 cm), and then evenly distributed down through the rest of the profile. Nearly 30 and 50 % of the entire pool of respectively N and C is in these top 25 cm forming a characteristic A-horizon, as a result of pedological processes. This also shows that decomposition happens slowly because the %C goes towards 0 already in 20 cm depth.

Conclusion

It can be concluded that there in the past never have been any large activity including a high content of organic matter at least not in the period creating the top 2 meters of the soil. Melting of the top permafrost in this area, will due to the very little carbon and nitrogen content down through the profile not lead to any significant GHG-emissions and the global warming potential is therefore very little.

An interesting indication seen from our data, specifically the grain size data, is a great sandstorm or sand that have been deposited continuously over a long time period which buried the old surface with such a thick a layer that a new surface have developed.

In a geographic spatial perspective, these data are very site specific and should in future studies only be used for up scaling or modeling in Flakkerhuk, East Disko.

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Climate Gradients at Flakkerhuk, Disko

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Abstract. To predict how changes in future climate and temperatures in particularly, can influence the permafrost it is necessary to understand how different climatic parameters influence each other at present climate. Apart from local influences is temperature controlled by an inland effect and the dry adiabatic lapse rate. The effect from these is analysed and an inland gradient was found. This could be used in permafrost modelling beyond the location where temperatures were obtained.

Keywords: *Inland gradient, lapse rate, climatology.*

Introduction

To examine the behaviour of permafrost it is necessary to examine and understand the local climate. Combined with the texture, thermal properties and water content, the energy available at the surface controls the thawing and freezing of the soil. Climate measurements are therefore important in the understanding of the development of the active layer depths. Due to time and logistical constraints this article presents climatic parameter data for a limited area and period in July 2011. The different climatic parameters that have been measured at the field site at Flakkerhuk, Disko Island (see appendix) includes; air temperature (T_{air}), relative humidity (RH) and wind speed in all 2m height, surface temperature (T_{surf}), soil temperature (T_{soil}) in two depths (in 5cm and 10cm at a wet spot and in 10cm at a dry spot), soil heat flux (G) at 1-2 cm depth, solar radiation and net radiation. The measurements have been conducted from July 7th – July 18th. From the 10th of July temperature, relative humidity and wind speed have been measured with a additional mast on a transect covering a distance of 3.5 km from the coast and inland. The most important climatic parameter for permafrost modelling is the temperature. Temperature is either affected by the other measured parameters or influences them. How temperature is affected and affects these parameters is examined. The irradiance varies north of the Arctic Circle during the year. In winter there is a period where the sun never rises above the horizon (November 29th – January 11th) and in summer there is a period where the sun never goes below the horizon (May 20th – July 22nd) (Bruun et al. 2006) (See also overall introduction for additional information). In summer, irradiance is therefore high while low in winter. Irradiance influences the radiation balance where the net radiation is the accumulated contribution of incoming and outgoing, long and short waved energy. The net radiance is therefore the total amount of energy available at a given location. Consequently the net radiation and soil heat flux are part of the energy balance of this location and thereby also influences temperature (Oke 1987). Apart from the measured parameters temperature is influenced by a number of other factors. There can be large local variation e.g. due to difference in surface type. On a larger scale the temperature at Flakkerhuk is affected by the large cold water body around Disko. The high specific heat capacity in water should make the coastal areas colder compared to further inland. Rouse, W.R. (1991) found that the wind direction had a major affect on the summer T_{air} at Hudson Bay. The T_{air} difference between onshore and offshore wind is on average 5 °C in the growing season. The effect is greatest the first 3 km after the wind crosses the coast, but can be seen 65 km inland. Under offshore winds there is only a small increase in temperature from the coast inland. The temperature inland is also affected by the lapse rate. Disko is a mountainous island and the temperatures vary due to differences in elevation.

The combined effect of these two was investigated at Flakkerhuk in the period from July 10th - 18th. It is investigated how the lapse rate and the inland effect each influence the mean air temperature. This allows extrapolation permafrost modelling beyond the site where the temperatures were obtained.

Methods

The radiation balance is given as;

$$Q^* = K\downarrow - K\uparrow + L\downarrow - L\uparrow$$

Q* is net radiation. K ↓ and K↑ is short waved, incoming and outgoing, radiation, where L↓ and L↑ is the long waved, incoming and outgoing, radiation (Oke 1987). Only Q* and K ↓ was measured at the stationary mast July 7th-18th. The rest of the parameters can be calculated. The outgoing short waved radiation is controlled by the surface albedo (K↑/K ↓). The albedo can be obtained from the Arctic Station in Godhavn. An average albedo for July was found to be 0,155. This corresponds well with the findings of others e.g. Dutton et al. (1991) that had albedo values ranging from 0.14-0.17 in July and Lafleur et al. (1997) who in July measured the albedo in mixed tundra to be just below 0.16. Changing surface conditions will affect the albedo. A wet surface will have a lower albedo than a dry surface. This can have an effect after rain and later in the summer period, when the surface gets drier. A simple way of expressing L↑ is given by the Stefan-Boltzmann Law (Oke 1987). It describes the radiant flux density (W/m2) emitted from a body.

$$L\uparrow = \sigma T_0^4$$

σ is the Stefan-Boltzmann constant ($5.67 \cdot 10^{-8}$ W/m^2K^4) and T_0 is the temperature in Kelvin. This equation presumes that the body is a perfect black body with an emissivity of 1. For simplicity it is assumed that the surface has an emissivity of 1. L↓ is found as the residual.

To determine the inland effect and lapse rate effect, two climatic stations that, among other parameters, measured the air temperature in two meters height have been used. One was permanently placed, close to the coast in a wet area, while the other was moved inland every 24th hour. By doing this it was possible to compare the air temperature at two locations at the same time and thereby get the effect of the lapse rate and inland movement. This was done for eight locations covering a distance of 3.5 km. Each location was randomly picked in a fairly uniform landscape, mostly northern willow (*Salix glauca*) and dwarf birch (*Betula nana*). However, the first location was at the beach ridge which did not have any vegetation.

Results and discussion

Figure 1 shows the variation in Q*, K↓, K↑, L↑and L↓. During the day net radiation is positive meaning there is an energy surplus coming from the atmosphere to the surface. Only a short period during the night has a negative net radiation. It shows a very clear diurnal cycle even though it never gets dark. This is due to low irradiance when the sun goes below the mountain ridge in the very late and early hours of the day. K↓ is directly affected by K↑ due to the albedo and therefore show the reversed signal of K↓. L↓ and L↑ do not vary as much as the other parameters during the day. The net contribution of the long waved parameters is usually negative. At night, when K↓ and K↑ are very small, there is an energy loss from the surface due to the negative net long-wave budget.

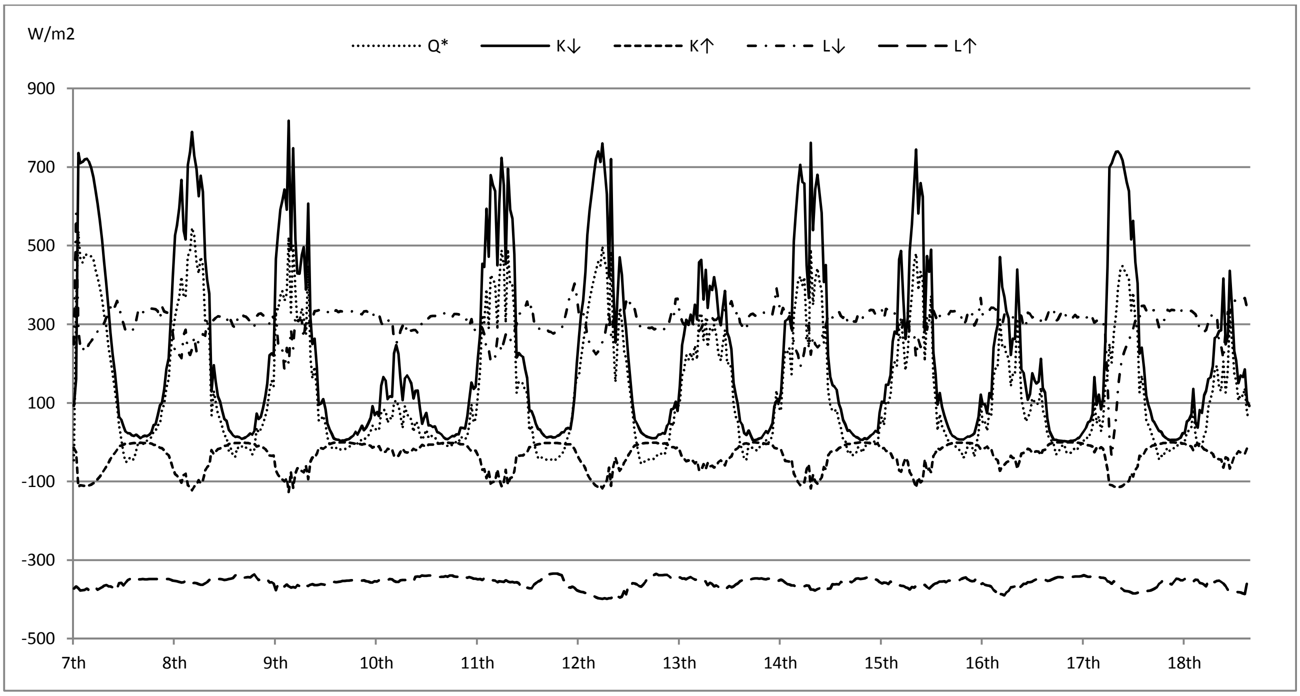

Figure 1: Net radiation, short-wave incoming and outgoing radiation, and long-wave incoming and outgoing radiation for the stationary mast July 7th-18th.

The relatively cloud free days have peak irradiance between 700-800 W/m^2, and a net radiation ranging from 400-550 W/m2. On days with cloudy weather e.g. July 10th, the surface receives less energy and has an 80 % lower net radiation. Q* and G are part of the energy balance of a given location. Figure 2 show how soil heat flux also follows a diurnal cycle.

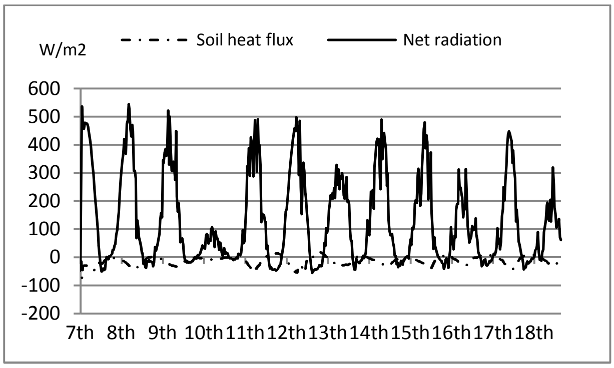

Figure 2: Net radiation & soil heat flux July 7th-18th

The diurnal cycle is reversed compared with net radiation. During the day the soil heat flux is negative, hence there is a transfer of energy from the surface into the soil and T_{soil} increase. During night there is a transport of energy from the soil to the surface and heat is emitted from the soil. The soil heat flux consequently influences the temperature variations in the soil. Figure 3 show the soil temperature in 5cm and 10cm depth in a wet soil and in 10 cm depth for a dry soil.

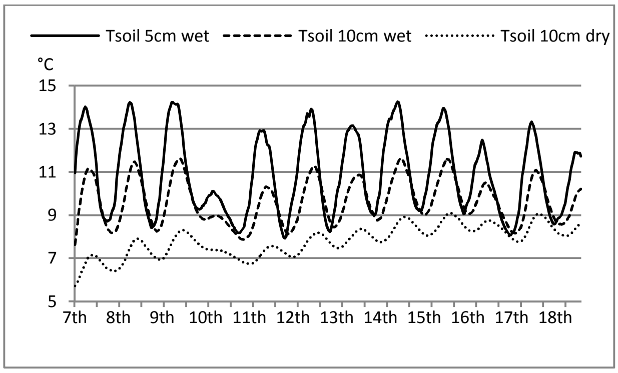

Figure 3: Soil temperature measured at the stationary mast July 7th-18th. Soil temperatures were measured in 5cm and 10cm at a wet spot and in 10cm at a dry spot.

There is a clear diurnal signal in all of the measurements. The day to day variations correspond very well to the variations in soil heat flux seen in Figure 2. Surprisingly, the biggest day to day variation is seen in the wet soil. The opposite should be expected because of the higher specific heat capacity of water. An explanation could be that the dry measurements were taken in a hummock with a thick layer of vegetation that could function as isolation. The wet soil did not have a thick layer of vegetation and is therefore more prone to variations in surface temperatures. As expected there is a bigger variation in 5cm than in 10cm. Figure 4 shows the daily T_{air} and net radiation.

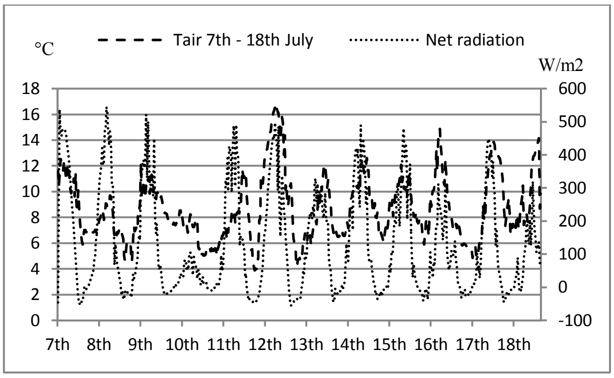

Figure 4: Air temperature and net radiation measured at the stationary mast July 7th-18th.

At July 10th, the net radiation as well as the temperature is very low. July 10th has the coldest daily mean air temperature in the measuring period. Overall there is a good correspondence between net radiation and air temperature. However it does not explain the relative low temperatures at July 8th when net radiation is as high as the rest of the days, nor does it explain the relatively high temperatures at July 16th where the net radiation is much lower than the average. In the first case higher temperatures are expected. Hence, there must be a transfer of energy away from the area, which causes the lowering of the air temperature. July 16th is expected to be colder than shown in the measurement, which means that there is a transfer of energy into the area. This could be explained by wind direction. If the wind direction is from the mountains the air would be influenced by the lapse rate and the coastal effect would not be so pronounced. This would make the air temperatures increase, as the air flowing into the area is warm and dry as it has travelled down the slope and thereby undergone change in pressure. Unfortunately wind direction data is not available from our field measurements. General air mass movements can also explain the differences seen these data. Warm air coming from south could explain the higher, than expected, temperatures the 16th. Cold air masses from north could explain the colder, than expected, temperatures the 8th (Oke 1987). A lower temperature due to greater evaporation can be disregarded, since the 7th was not rainy. Rain would make the surface wetter and therefore support a greater evaporation, which would lower the temperature. Figure 5 shows that air temperature and relative humidity have reversed covariation. Warm days have a low relative humidity whereas cold days have a high relative humidity. At night the relative humidity reaches 100 %, which is the dew point. July 10th was a rainy day, which explains the relative humidity of 90-100 % during the day.

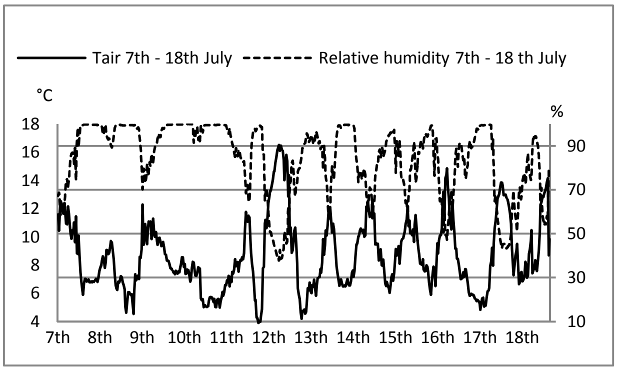

Figure 5: Reversed covariation between air temperature and relative humidity. Measured at the stationary mast July 7th-18th.

Figure 6 shows T_{air} and T_{surf}. Generally, the surface temperatures are higher than air temperatures during the day whereas the difference equals out during the night. Again July 10th and July 16th stands out. These days differ from the rest by having almost the same surface and air temperatures. This highlights that 'external' influences other than irradiance or net radiation affects the temperatures locally e.g. the general air mass movement.

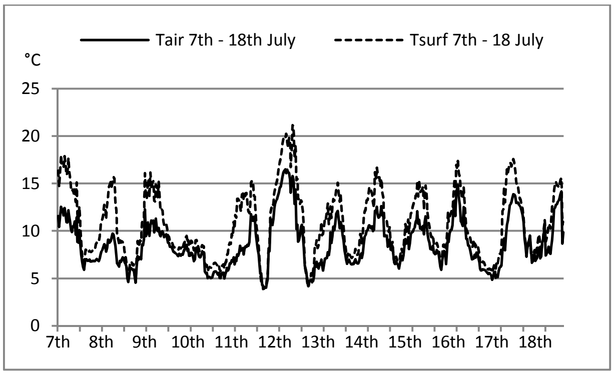

Figure 6: Air temperatures and surface temperatures. Measured at the stationary mast July 7th-18th.

Topography influence

From July 10th-18th a secondary climate mast measured $T_{air,}$, relative humidity and wind speed in 2m height. Higher temperatures inland were expected because of the inland effect. For simplicity reasons mean daily temperatures have been used in Figure 7.

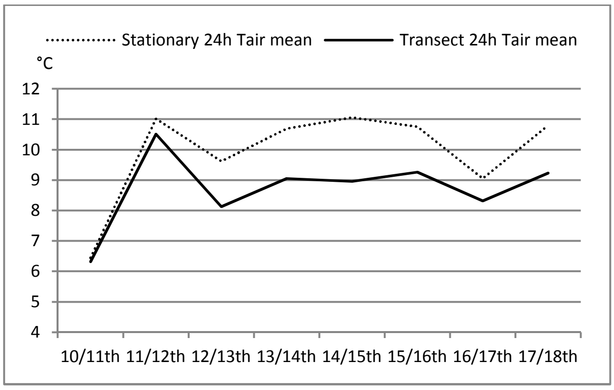

Figure 7: Comparison of mean air temperatures of the stationary and portable masts, July 10th-18th.

The differences can be controlled by the inland effect, the lapse rate or by several other climatic influences. In the following the influence from the different effects will be discussed. As the figure show there is an increase in temperature difference until the 14th/15th. This is most likely due to the inland effect. From here the difference decreases until the 16th/17th, where after it increases again. The increase in mean air temperature between the stationary mast and the last location 3.5 km inland is 1.56 °C. To get an understanding of the forces driving this trend, the mean temperatures for respectively the warmest (09.00-15.00) and coldest (21.00-03.00) period of the day are shown in Figure 8 and Figure 9. The relation between the date, elevation and distance to the coast can be seen in Table.

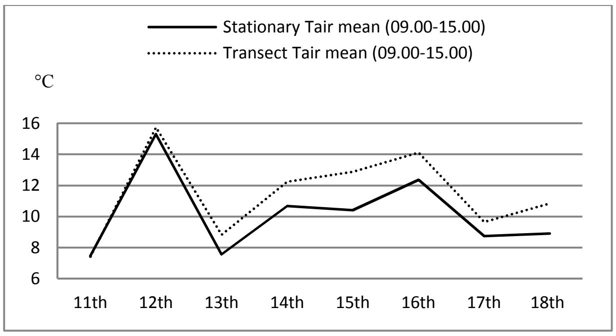

Figure 8: Comparison of mean air temperature at the warmest period of the day (09.00-15.00). Measured at the stationary and portable masts, July 10th-18th.

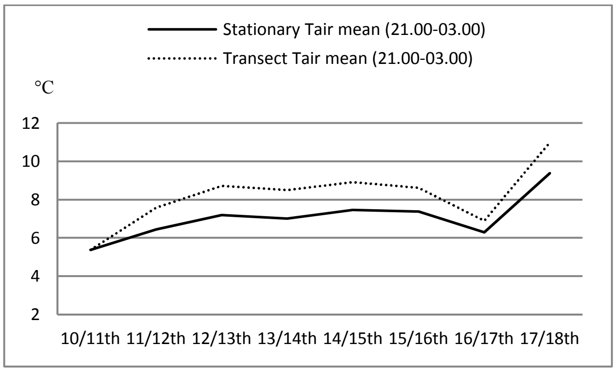

Figure 9: Comparison of mean air temperature at the coldest period of the day(21.00-03.00). Measured at the stationary and portable masts, July 10th-18th.

The figures show that even though there are some differences in the patterns, the same trend is seen. The biggest differences are seen at warmest period of the day. The biggest difference on 2.5 °C is seen the 14th/15th. The smallest differences occur during night. The differences during night do not vary as must as at day. For both periods there is an increase in temperature difference until the 14th/15th, after which the differences in temperature decreases until 16th/17th. Again the period ends with an increase. This suggests that there is an overall inland effect which is not controlled by specific events occurring only during days or nights (See Table 1 for the exact temperature difference values).

Table 1 Temperature differences between the stationary mast and the transect, July 10th-18th.

	Difference in °C between the masts				
Date	**24h mean**	**09.00-15.00**	**21.00-03.00**	**m.s.l**	**Distance from coast (m)**
10/11th	0,1	0,0	0,0	0	0
11/12th	0,5	0,4	1,1	2	163
12/13th	1,5	1,3	1,5	6	408
13/14th	1,6	1,6	1,5	38	1022
14/15th	2,1	2,5	1,4	54	1443
15/16th	1,5	1,7	1,2	53	1899
16/17th	0,7	0,9	0,6	78	2445
17/18th	1,6	1,9	1,6	116	3388

Relative humidity varies between each location as seen in Figure 10. The relative humidity varies a little, at daytime, between the days during the period, but has equal values within the same days for the two masts. During the day of the 14th there is a difference in daytime relative humidity between the stationary mast and the portable mast, with a higher value at the stationary mast.

This could be explained by temperature difference that day, which was the highest in the period.

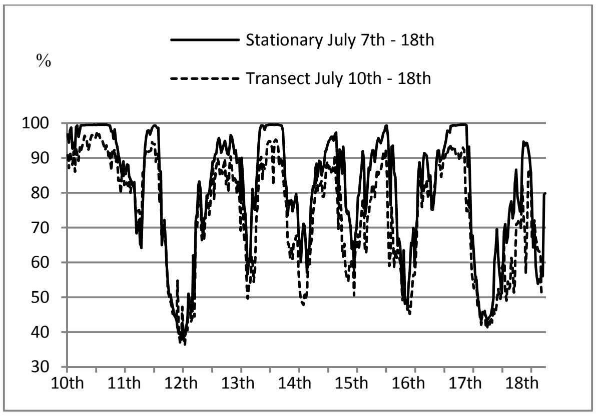

Figure 10: Comparison of relative humidity measured at the stationary mast and along the transect July 10th-18th.

The higher temperature reached during the day at the portable mast would lower the relative humidity relatively more than it would be the case at the stationary, given that the air masses had the same water content. Figure 11 show the actual vapour pressure of the air and thereby the actual amount of water in the air.

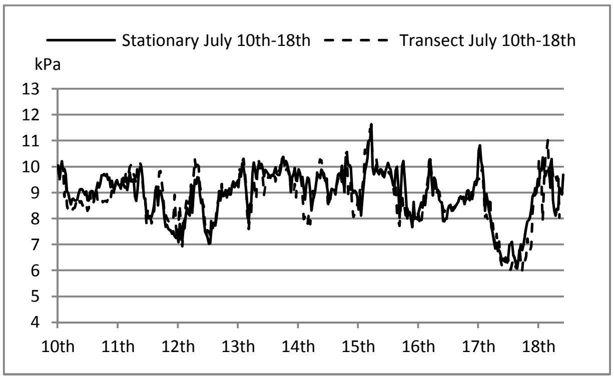

Figure 11: Actual vapour pressure measured at the stationary mast and along the transect July 10th-18th.

The figure show small variations in the actual vapour pressure between the stationary mast and along the transect. This shows that the higher relative humidity at the stationary mast can be explained by lower temperatures.

Another explanation for the lower temperatures at the stationary mast could be the surface type. The stationary mast was placed in a fen area with small water puddles around it. During the day the temperature would be lower due to more energy going to latent heat than sensible heat, compared with a dry site. At the same time there is enough wind to exchange the air masses above the surface so that the relative humidity does not increase at the wet area. Figure 12 show the wind speed measured at the two masts. Every day the portable mast was moved inland, but no difference between the masts can be observed. The same trend and magnitude was seen for each location.

However, the surface type explanation can only explain a temperature difference between two locations and not the increase in difference inland, which is seen.

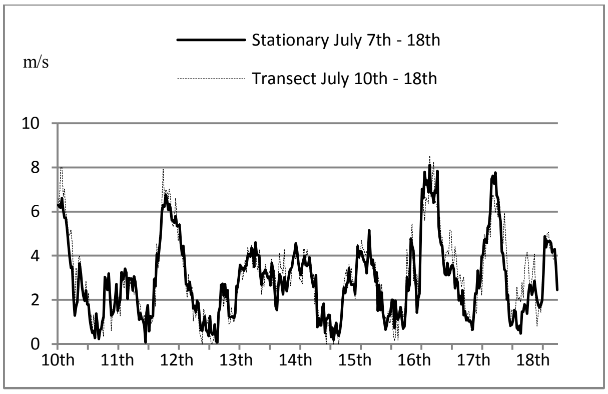

Figure 12: Comparison of wind speed measured at the stationary mast and along the transect July 10th-18th.

The temperature is also influenced by the lapse rate. Figure 13 show the elevation of the points in which the climate station has been placed along a transect inland. There is an increase in elevation the further inland the climate station was moved. Along this transect it is examined how the air temperature is influenced when moving inland and upwards in altitude. To see the relationship between locations and date see table 1.

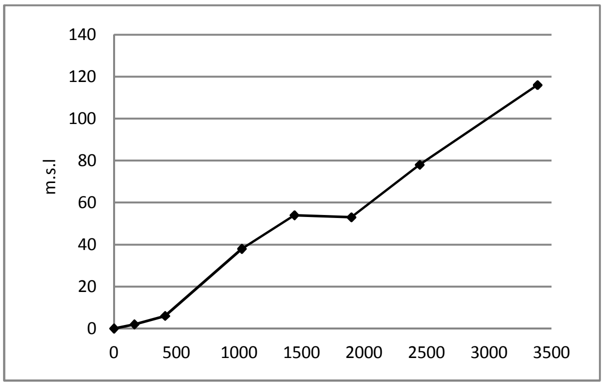

Figure 13: Elevation and distance to coast at the 8 locations.

It is assumed that the air temperature in an air mass decreases dry adiabatic in the order of 0.65° C per 100 m (Dewalle 2008, Jensen 2009). Figure 14 show the lapse rate effect at each of the eight points. At the beach the lapse rate is negligible, while the temperature change increases when moving inland as elevation increases.

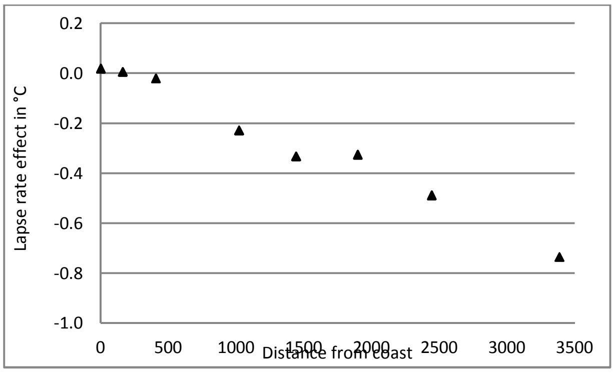

Figure 14: Lapse rate effect at the 8 locations when portable mast was moved inland July 10th-18th.

By knowing the lapse rates effect on the mean air temperature, the pure inland effect can be obtained. This can be done by adding the lapse rate for each location to the site-specific mean air temperature. The result can be seen in Figure 15.

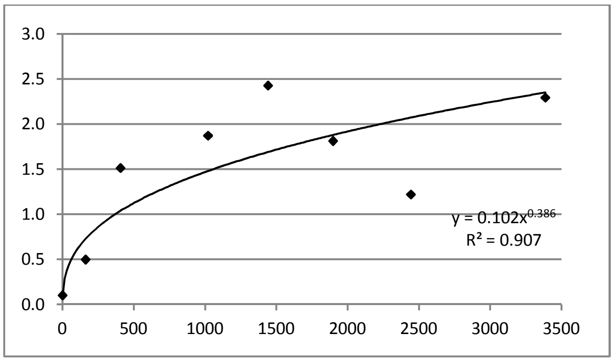

Figure 15: The pure inland effect disregarding the lapse rate effect over a distance of 3.5 km.

Overall an inland trend can be seen, from the first point located at the beach ridge until the last point placed 3.5km inland. The inland gradient has been calculated as seen in Figure 14. The inland temperature increase is expressed by an exponential regression, which gives a good fit with a R^2-value of 0.907. It shows how the effect is greater close to the coast and decrease further inland. The temperature difference at the location 2500m inland corresponding to the dates 16th/17th does not seem to be explained be the coastal-inland effect. If temperatures this were affected by wind, and the transport of warm and cold air masses, the inland trend would not be observed at days where the surrounding climate played a big role. If the reasons are more site-specific, surface type and wetness could have an effect. A completely dry surface would be warmer since no energy would be used for evaporation. Haugen et al. (1980) found a different coastal-inland relationship. They found that the temperature increased from 5 to 12 °C 350 km inland. Using the regression found in this research gives an increase of 14.1 °C 350 km inland. Comparing the results with those of Rouse (1991) also show an overestimation of the inland effect further inland. He found that the temperature was 4.4 °C higher 65 km inland, whereas we get 7.35 °C. The results are not directly comparable. This research only includes measurements from locations 3.5 km inland, whereas Haugen et al. (1980) includes measurements from locations 350 km inland. Finally it should be recognised that the study area is too small, and the sample time too short, to produce really reliable results.

What we have seen can be small daily variation controlled by a number of different factors.

Conclusions

An inland trend in the magnitude of 1.56 °C can be seen from the coast 3.5 km inland. A relationship between temperature and distance to the coast has been found, but overestimates the inland effect compared to other findings on the subject. Determining the inland effect and the lapse rate effect allow for extrapolation of the temperature beyond the site where the measurements were conducted. This can be used in permafrost modelling where temperature is the most important factor.

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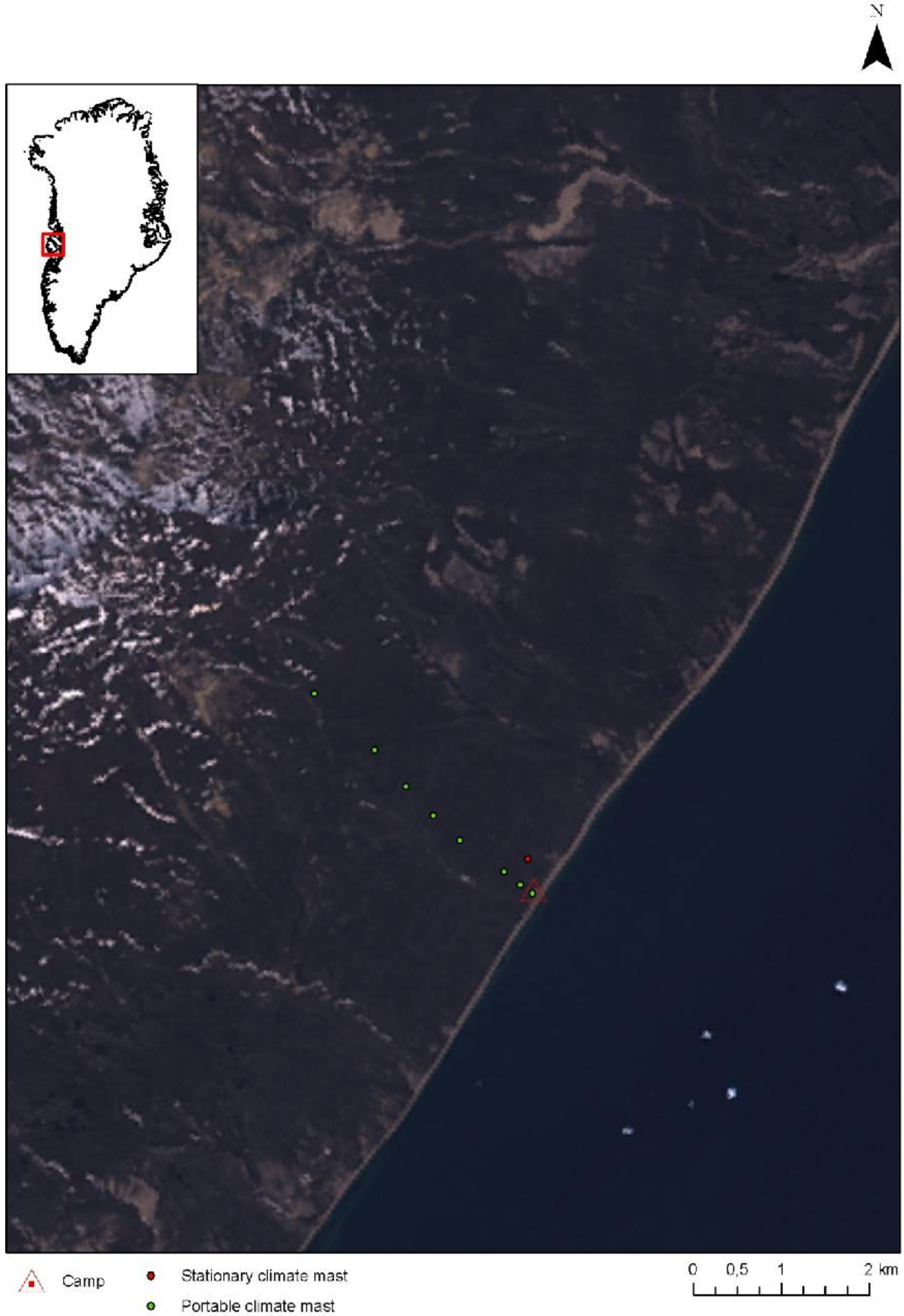
Appendix


N
Camp
Stationary climate mast
Portable climate mast
0 0,5 1 2 km

Methane fluxes measured at Flakkerhuk, Disko Island (West Greenland) at different vegetation specific sites.

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Abstract. CH_4 emissions were measured at two wet sites and two dry sites during two weeks (July 2011) in Flakkerhuk, Disko, Greenland (69°N). Sites were chosen based on different vegetation cover. The wet sites released on average 4 mg $CH_4/m^2/d$ while the dry sites revealed an uptake of 1.3 mg $CH_4/m^2/d$. No correlation between CH_4 and soil water content and soil temperature were found. A better understanding of soil structure, pH, amount and quality of organic matter, vegetation and bacterial composition might have verified the spatial differences in net CH_4 emissions.

Keywords: *CH_4, production, consumption, Greenland.*

Introduction

Methane (CH_4) is one of the most important greenhouses gasses, beside carbon dioxide (CO_2) and water vapour in respect to climate change. CH_4 has a warming potential which is 25 times higher than CO_2 on a 100 year scale (IPCC, 2007). About 30 % of the global emission is from natural sources where wetlands account for 70% of this emission (Megonigal, 2004; Bonan, 2008). In the Arctic the ecosystems are very much influenced by the presence of permafrost. The existence of permafrost preserves the labile organic matter from decomposition (Riley et al., 2011).

However, in the past centuries increase in global temperature has resulted in thawing of the permafrost and made the organic matter accessible for decomposition (Riley et al., 2011). Especially in low-lying areas and depressions the thawing of permafrost can create anaerobic conditions which favour the production of CH_4, resulting in a positive feedback to climate change. The thawing of permafrost in drier areas will also create a positive feedback to climate change with enhanced CO_2 emissions from decomposition. Northern wetlands are today estimated to emit around 10-40 Tg CH_4 yr^{-1} (IPCC, 2007).

Atmospheric CH_4 can however also be oxidized in some ecosystems that have a drier environment. These biological sinks are responsible for 6 % of global CH_4 consumptions (Roger and Le Mer, 2001).

CH_4 is produced in anaerobic environments below the water table by methanogenic bacteria, archaea, as the final step in the series of processes that degrade organic matter. The CH_4 is either emitted directly to the atmosphere or oxidized to CO_2 by methanotrophic bacteria in the aerated part of the soil above the water table (Parmentier et al., 2011).

CH_4 produced in the soil can be emitted into the atmosphere by several pathways and is therefore more unpredictable than CO_2. One way is simple diffusion through the soil profile, which is 10^4 times slower in water than air (Hanson et al, 1996). A second pathway is through ebullition, meaning sufficient gas will be produced as small bubbles in the saturated layer and force their way to the surface. The third way is through the continuous air space plant tissue known as aerenchyma. Those are found in vascular plants, which has adapted to flooded environments (Smith et al., 2003). This will provide a direct pathway to the atmosphere, bypassing aerated zones in the soil where CH_4 would normally be oxidized

(Parmentier et al., 2011). Sedges or grasses such as *Carex L* and cotton grass, *Eriophorum vaginatum* are some of the plants, which have this pipeline effect (Parmentier et al., 2011). However, the abovementioned aerenchyma can also increase availability of oxygen in the rhizophere. This will therefore stimulate the methanotrophic bacteria living in aerobic environments, to oxidize CH_4 (Ström et al., 2005).

The net balance of CH_4 fluxes from a given ecosystem is therefore firstly controlled by the soil water content, which determines whether CH_4 production or CH_4 oxidation is the dominant process, yet also the balance of CH_4 production and uptake within the same environment. Both processes are however influenced by soil temperature and pH (Schneider et al., 2009).

In general it can be summarized that level of water table, soil water content, type of vegetation and temperature are some of the most important parameters which influence CH_4 production and oxidation (Hanson et al., 1996).

The aim of this paper is to analyse the spatial and temporal estimate of CH_4 net emissions from different vegetation types at Flakkerhuk, Disko, Greenland.

Method

Study area

The study area is situated at Flakkerhuk, Disko Island, Greenland (69°N, 53°W) about 50 kilometres northwest of the town of Ilulissat. The area is placed in a high arctic environment. Flakkerhuk is primarily flat with a gentle elevation increase inland often characterized as lifted marine terraces. Within short distances some areas had the presence of high water table indicating permanent wetland condition, which were further empathized by the vegetation type found in these areas. Other areas had no visible water table and the vegetation type in these area indicated drier sites. Four sites were selected and named based on dominant vegetation cover or composition, hummocky fen (HF), arctic cotton (AC), dwarf willow (DW) and dwarf birch (DB). Where HF and AC represents wetlands and DW and DB represent dry sites. All four sites were located on the youngest marine terrace, which is explained in more detail earlier in this report.

HF was situated at 22W 0461322 7717546 and measured approximately 20 x 20 metres with approximately 40 % being hummocks. The rest is exposed water table. The hummocks had a very diverse vegetation composition. Different types of willow were present such as *Salix arctophila* and *Salix arctica.* Herbaceous plant such as *Pyrola grandiflora,* was also present. Grasses such as *Carex L.* and *Deschampsia alpina* were also dominating the hummocks. For some of the hummocks mosses as *Sphagnum* was also present. The chambers with exposed water table were primarily dominated by *Carex L* and *Sphagnum.*

AC is situated at 22W 0461341 7717470 and measured app. 10 x 7 metres. The area had permanent high water table. Several stones and pebbles were visible on the ground. The area was dominated by grasses such as *Eriophorum vaginatum* and *Arctophila fulva*. Moss as *Sphagnum* was also dominating the area.

DW is situated at 22W 0461308 7717470 and represented dry site conditions. *Salix arctica, Salix glauca* and bare soil were present in a mosaic pattern and big stones made the area rugged. DB covered a much more regular plane. This site was dominated by *Betula nana* yet also the Arctic bell-heather *Cassiope tetragona* and Mountain Crowberry *Empetrum hermaphroditum*. DB coordinate is 22W 0461312 7717377.

Instrumentation

For measuring CH_4 concentrations a mobile LGR-DLT100 (Los Gatos Research, http://www.lgrinc.com) was used. From this instrument two Teflon tubes (5 metres long and with an air volume of 141 cm^3) were connected to a mobile lid defined as an outlet and inlet to the system. The Teflon tubes were installed to minimize gas diffusion with the surroundings.

Furthermore two thermometers were installed inside the lid measuring the temperature inside and outside the chamber. This enabled analysis of temperature dependent CH_4 fluxes. Board walks were used to minimize disturbances of the soil. Once the lid was placed on the chamber, the system would be closed and disturbance from ambient atmosphere therefore could be negligible. Furthermore, to diminish any heating effect from the sun a shadow blanket was placed over lid and chamber during measurements. To ensure circulation of the air inside the chamber a fan was installed under the lid. In order to analyse the effect of CO_2 a LICOR 840 was installed as well.

Procedure

Cylindrical dark chambers with a diameter of 31 cm were placed over the specific vegetation type dominant at the different sites and inserted app. 15-20 cm into the soil. Several replicates were placed within each site. Thus, HF had 20 replicates, 10 replicates around the hummocks (H) and 10 on the standing water table (WT), AC had 10 replicates, DW had five replicates and DB had 10 replicates, five of them around the vegetation and five were placed at the site, but where the A-horizon had been removed (DB no A) thus manipulated conditions.

Measurements were conducted over two weeks in July 2011. Each site was measured four times hence four campaigns are obtained, except from DB, which only got measured 3 times. At HF and DB furthermore a 24 hour campaign was conducted. This enables analysis on diurnal variability. One measurement took app. 15 min. The machine was designed to record the CH_4 concentration in unit ppm (parts per million) every 10 seconds. While running the measurement the lid was placed on the chamber. The first three minutes were a waiting period where conditions stabilized. Over the following ten minutes the values of CO_2 and CH_4 concentrations within the chamber were calculated and a flux was given. After removing the lid, five replicates of water content was measured around the chamber with a handheld Theta Probe, Soil moisture Sensors (ML2x Delta-T Devices ltd., Cambridge, UK). This sensor responds to changes in the apparent dielectrical constant. These changes are converted into a DC, voltage, which is proportional to soil water content according to Gaskin and Miller (1996). Soil temperature in 5 cm depth was measured five times around the chamber with a hand held thermometer.

The fluxes of CH_4 and CO_2 were calculated based on the equation *(1)*

$$Fc = \frac{\left(\left(\Delta C/\Delta t\right) * V * P\right)}{\left(R * T * A\right)} \qquad (1)$$

Where $\Delta C/\Delta t$ is the concentration gradient of either CH_4 or CO_2 over time, V is volume of the chamber, P is the atmospheric pressure, R is the gas constant, T is temperature inside chamber and A is the area of the chamber, and units were chosen to give the actual flux in $\mu mol/m^2/s$.

Results and Discussion

Emission and uptake

Throughout the field work a total of 300 chamber measurements were carried out. Most measurements had a steady increase in CO_2 concentration, as expected due to the ongoing soil respiration (Han et al., 2007). This CO_2 increase was therefore a good indicator for validation of the measurement. In Figure 1 a classic example of steady increase of both CH_4 and CO_2 from one of the wet sites can be seen. CH_4 concentrations increases from around 1.87 ppm to 2.05 ppm over the ten minute measuring period, and for the same period CO_2 increases from less than 400 ppm to app. 500 ppm.

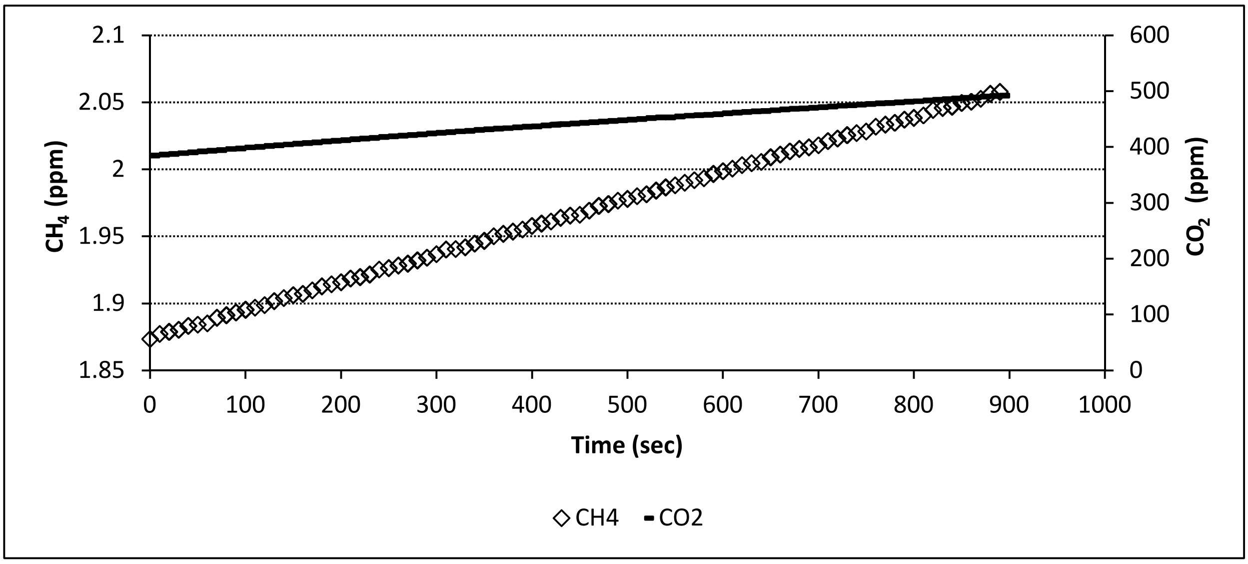

Figure 1. A classic example of CH_4 and CO_2 increase at a wet site. Both concentrations over time are given in ppm. This example is taken from chamber 3 at AC.

In Figure 2 is given a classic example of the CH_4 and CO_2 development at a dry site. Here respiration rates (CO_2) increases from below 400 ppm to almost 800 ppm. Contrary to the former graph, a CH_4 uptake is now seen. CH_4 concentrations starts around the atmospheric concentrations level which is around 1.8 ppm and through the next 10 minutes it is clear that CH_4 is being consumed, resulting in a CH_4 concentration around 1.76 ppm after end measurement. Values are small yet nonetheless it is clear whether a CH_4 production or consumption is taken place at a wet site or a dry site respectively.

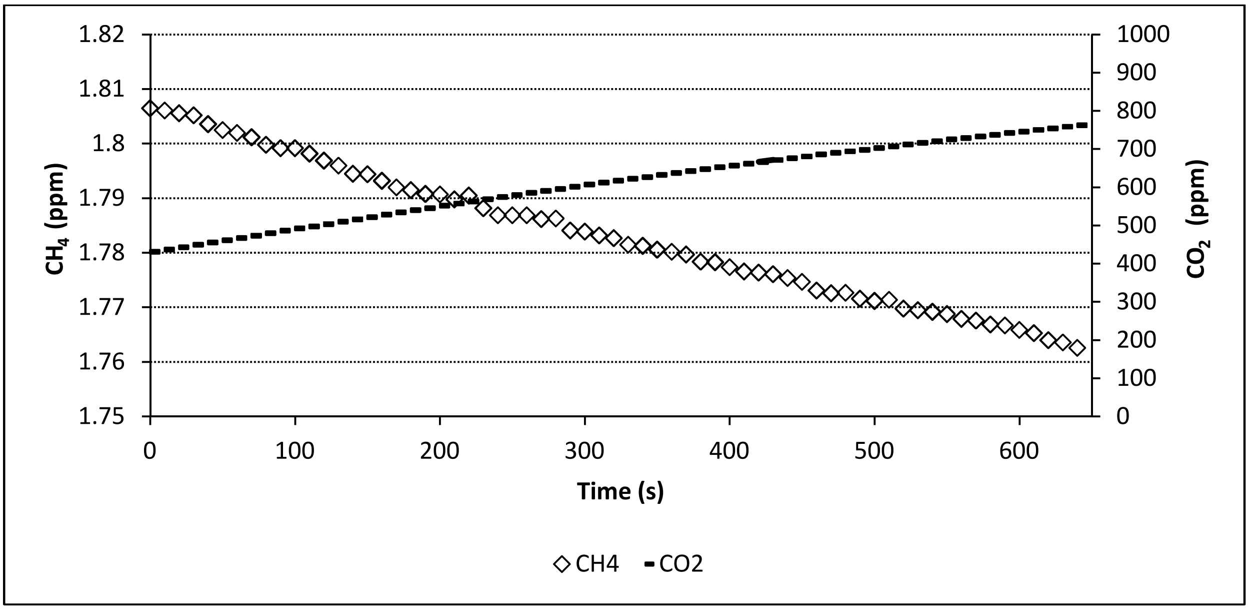

Figure 2. A classic example of CH_4 and CO_2 concentrations at a dry site. Both concentrations over time are given in ppm. This is measured at chamber 5 at DW.

Some measurements are considered doubtful and therefore removed from further analysis. This can be due to wrong chamber installation, technical or operational mistakes. From figure 3 CH_4 is varying between 1.88 and 1.888 ppm throughout the measurement period. These values are very small and show no clear trend, neither a CH_4 production nor consumption. This is very different compared to the general pattern of measurement. Furthermore CO_2 shows a decreasing trend, which is very different compared to the general pattern of measurement, which was conducted in this study. This specific example could be due to chamber conditions such as leakages, but also some chambers being placed right on top of big stones, can cause disturbances in the CH_4 and CO_2 concentrations. Those measurements are therefore removed from further analysis since these mistakes were measured repeatedly at the same locations throughout all the campaigns. A total of eight measurements at AC were removed.

The average CH_4 fluxes from the four sites and further sub division of HC into 'WT' and 'H' and DB into 'DB' and 'DB no A' are shown in Figure 4.

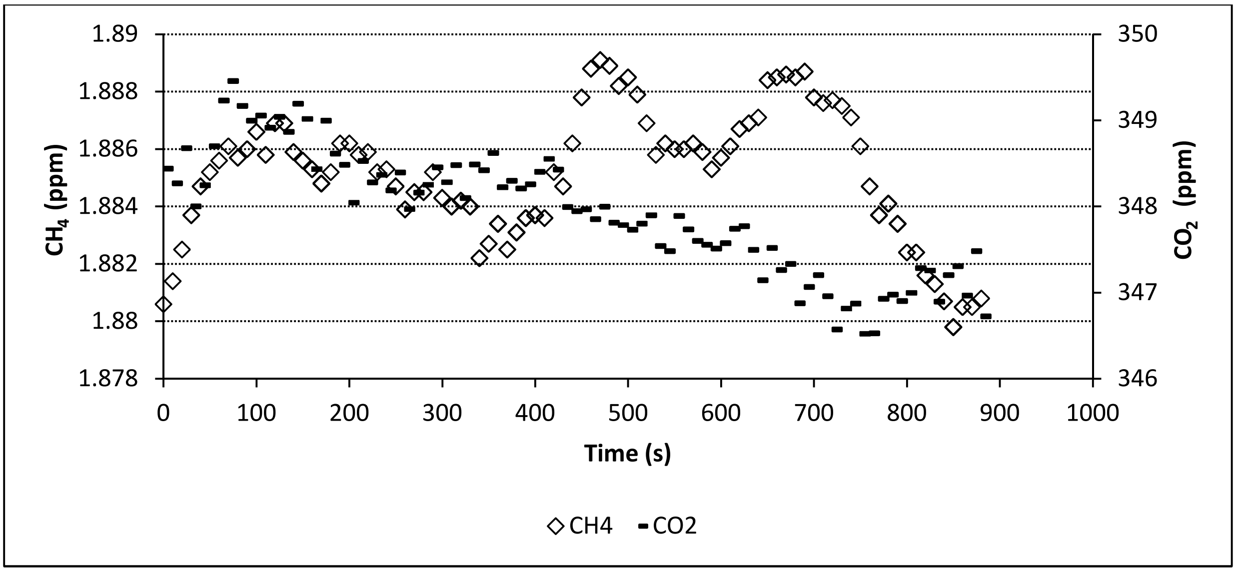

Figure 1. Example of measurement failure. Indication of chamber leakage, due to the decreasing trend of CO_2, and too many variations of the CH_4 flux. This specific example is from chamber 6 at AC.

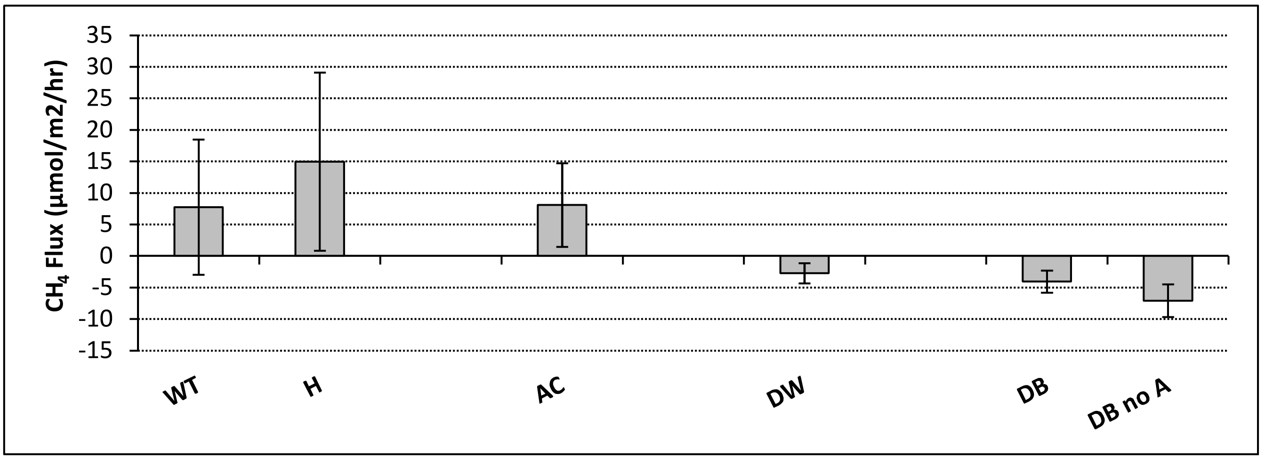

Figure 2. Average CH_4 fluxes measured at each site over the 2 week period. The CH_4 flux is given in µmol/m2/hr. The wet sites indicate a CH_4 production and the dry sites shows a CH_4 consumption.

The maximum CH_4 release was measured at HF with an average flux of 10.6 µmol/m^2/hr, where H is responsible for 15 µmol/m^2/hr and WT with an average of 7.8 µmol/m^2/hr. At AC the average production is 8 µmol/m^2/hr. Large standard deviations are seen for the wetlands indicating that these ecosystems are very sensitive to disturbances. Yet this has been tried to be reduced to its minimum e.g. by using board walks before placing the lid on the chamber.

The dry sites show downward fluxes in Figure 4 represented by negative values thus indicating a CH_4 uptake. On average DW shows an uptake of 2.7 µmol/m^2/hr which is lower than at DB where the uptake with and without A-horizon is 4.1 µmol/m^2/hr and 7.1 µmol/m^2/hr respectively. Measurements conducted at the dry sites showed small value variations through out the two weeks, also seen by the low standard deviation in Figure 4. This indicates rather stable conditions e.g. was not disturbed by the activity that took place around the chamber while measuring. Due to the short period of measurements very small changes in the climatic parameters, such as soil temperature and soil moisture content are seen, and the results are therefore to be seen as a snapshot for the arctic midsummer conditions.

Wet sites

The fluxes at wet sites vary remarkably as presented in Figure 4. The hummocks show to emit app. twice as much as AC namely 15 µmol/m^2/h compared to 8 µmol/m^2/h. The main reasons for this, is probably due to differences in vegetation composition. Especially the hummocks showed significant plant diversity and a thicker peat layer. Furthermore other studies have verified that CH_4 emissions are lower at areas covered by mosses such as *Spaghnum*, since *Spaghnum* is a habitat for methanotrophic bacteria (Parmentier et al., 2011). Larmola et al. (2010) concluded that the oxidation potential was highest when the water table was at the moss layer. This was the case at AC, which might explain the lower emissions of CH_4 compared with HF.

Another explanation for the flux variations can be found in the interval distribution of the measurement as presented in Figure 5. It shows the distribution of measured fluxes at HF, from hummock (a) and water table (b) respectively and at AC (c). Common for the wetlands in this study show that most fluxes lie within the 0-10 µmol/m^2/hr interval. Fluxes above 25 µmol/m^2/hr are only measured at HF indicating that site specific parameters control the fluxes as described above. More of the vegetated hummocks have higher fluxes ranging from 30-60 µmol/m^2/hr yet only few measurements are in the very high end of the range and these outliers are to be analyzed carefully.

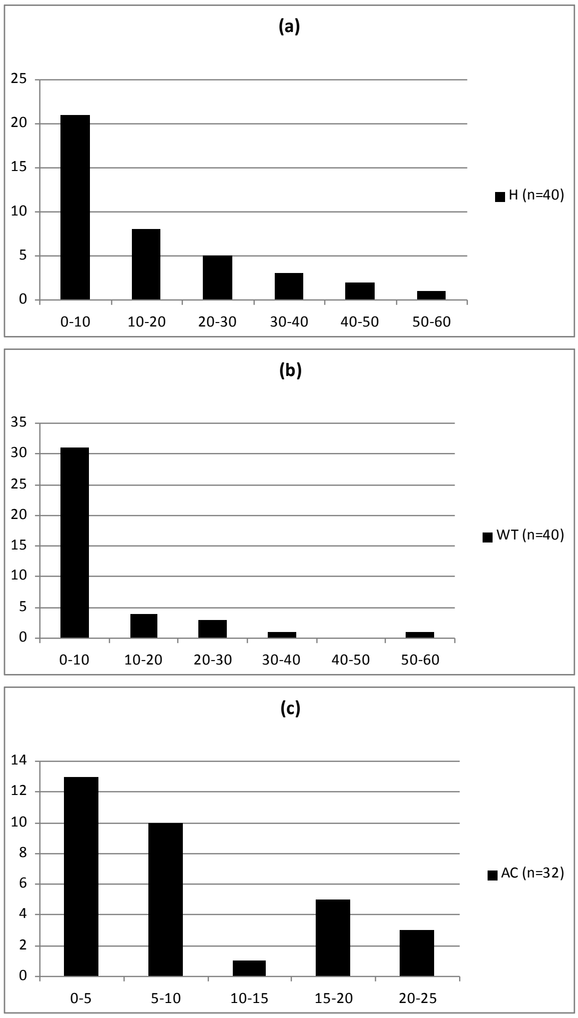

Figure 5. Distribution of chamber measurements. Intervals are given as CH_4 fluxes, µmol/m^2/hr. HF distribution is divided into hummocks (a) and water table (b). (c) is distribution at AC.

An example of such an outlier is a measurement at a hummock of 58.4 µmol/m^2/hr.

Ebullition

Outlying values can be caused by ebullition where CH_4 escapes as air bubbles and not diffusion driven.

In the wetlands most measurements showed a linear increase in concentration hence indicating simple diffusion through the soil profile or transport along roots and stems. Ebullition can cause a fast increase in CH_4 concentration when the bubbles reach the surface. Ebullition is a natural phenomenon, but can also be caused by disturbances of the soil. A few times the influence of ebullition occurred as presented in Figure 6.

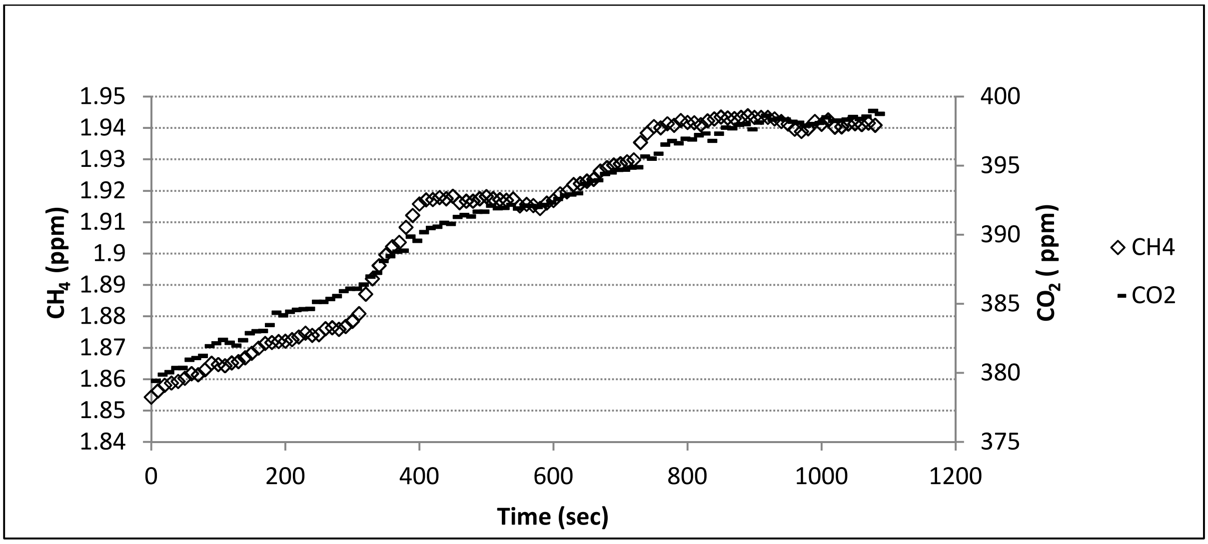

Figure 6. Example of a CH_4 bubble detected through the measuring period. This was measured at HF, chamber 14.

As it can be seen in Figure 6 a CH_4 bubble occurred at time 300 seconds where a steady increase is observable. Again at 700 seconds a small ebullition is also registered. This is further confirmed with the small increase of CO_2 which was detected at the same time. Such bubbles can affect the fluxes resulting in an enhanced flux which is higher than the average values.

Dry sites

Hardly any variations in soil water content are seen throughout the measuring period as presented in Figure 7. However from the Figure it is seen that DB is the driest site with water content less than 30 %, followed by DW with water content less than 40 %. At HF and AC the water content is listed as app. 100 %. This is a fairly incorrect value and more likely due to the setup of the Theta Probe indicating fully water saturation rather than the water content being actually 100 %. Soil water content is an important parameter controlling type of emissions either as CH_4 or CO_2. The more aerobic conditions the more the CH_4 will be oxidized due to microbial consumption and eventually emitted as CO_2. The CH_4 is consumed by methanotrophic bacteria which according to Bedard and Knowles (1999) favours aerobic environments. The methanotrophs consume atmospheric CH_4 as source of carbon and energy (Bedard and Knowles, 1999). In general DW has a higher water content, which might reduce the activity of the methanotrophic bacteria. DB being the drier of the two dry sites has the highest CH_4 consumption with on average 5.6 µmol/m^2/hr and DW 2.7 µmol/m^2/hr.

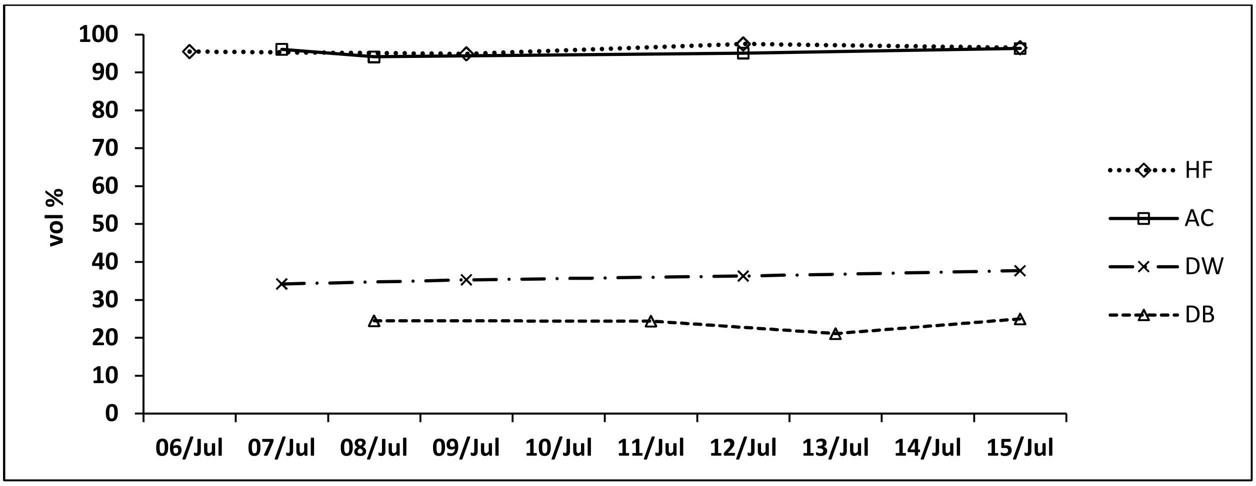

Figure 7. Water content measured at each four sites. With DB having the lowest water content with less than 30%, followed by DW which had a water content below 40%. The two wet sites both had water content more or less of 100%.

Manipulated site

In respect to dominant drivers of CH_4 uptake the manipulated measurements at DB where the A-horizon was removed showed interesting findings. These chambers have a higher uptake with a value of 7.1 µmol/m^2/hr as compared to 4.1 µmol/m^2/hr for chambers with the A-horizon. A similar finding is presented by Holmes et al. (1999). When investigating three different soil types in Denmark, Brazil and USA. That study reported the greatest oxidation activity of atmospheric CH_4 to take place in the mineral soil below the organic horizon (Holmes et al., 1999). It can be argued that the higher CH_4 uptake under the A-horizon is due to the composition of the methanotrophs. Another parameter influencing the CH_4 uptake is the temperature. Over four days of measurement an increase in CH_4 uptake from chambers without A-horizon is observed as seen in Figure 8. The uptake increases from app. 6±3 µmol/m^2/hr to app. 8±2 µmol/m^2/hr. Despite the short period of measurement the removal of the A-horizon might have caused the air temperature to have an enhanced effect on the soil processes below the surface as this got exposed to the ambient air. The activity of the methanotrophs under the A-horizon can thus be explained by Q_{10} factor with increased activity due to increased temperature. Hence, explaining the higher CH_4 uptake from chambers without the organic horizon. A further detailed analysis of the bacteria composition in different depths throughout the soil profile could be of interest concerning flux dependencies.

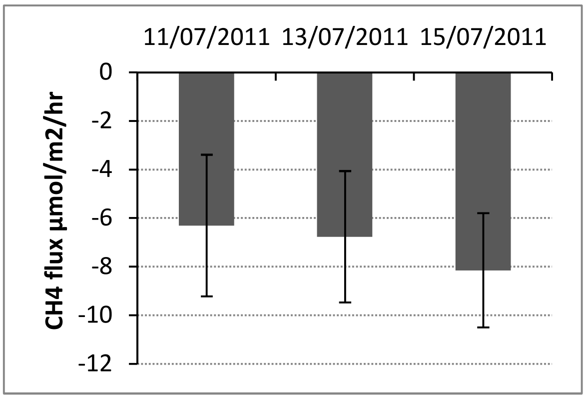

Figure 8. CH_4 fluxes at DB without A-horizon measured over four days in July 2011.

Lack of correlations

According to already well known and well documented theory, microbial activity rates vary with respect to soil temperature and soil water content (White et al., 2004, Bekku et al., 2004, Post et al, 1982, Roulet et al., 1992). However, no correlations between CH_4 and these parameters are found in this study.

Correlations were tested by using linear regression. At the wet sites CH_4 and temperature showed $R^2 = 0.0044$, $p = 0.48$. At the dry sites linear regression with temperature and CH_4 is $R^2 = 0.0237$, $p = 0.38$. Correlations between CH_4 and soil water content were just as poor from the wet sites ($R^2 = 0.018$, $p = 0.65$) and in the dry sites ($R^2 = 0.082$, $p = 0.10$).

Over the two weeks of measurement the temperature changes were too small to have an impact on the emissions. The soil temperature in 5 cm changed only a few degrees between the campaigns. The lack of correlation between CH_4 emissions and water content is consistent with issues discussed by Rinne et al. (2007). They argue that a poor correlation between CH_4 emission and soil water content is due to the fact that there are several parameters influencing the rate of emission. Therefore soil water content only becomes the dominating parameter when the water table drops to a certain depth causing more aerobic conditions hence reducing the emissions (Rinne et al., 2007). Correlations are expectedly enabled if measurements are performed continuously over a longer period that is, right after snow melt, at midsummer and at the end of the growing season.

24 hour measurement at HF and DB

24 hour measurements were conducted on three chambers at the HF site from 10 am to noon the following day. The results are seen in Figure 9 together with the mean soil temperature in 5 cm. The temperature varied between 7.9 °C and 10.7 °C with the maximum at 4 pm.

The CH_4 fluxes from all three chambers are very stable over the measuring period and no correlation to the soil temperature can be found. The lack of correlation between flux and soil temperature can be explained by the limited change in temperature throughout the day and it is indicated that the CH_4 production takes place deeper in the profile where the diurnal temperature change is even smaller. In the study of Rinne et al. (2007) from a Finnish fen the best correlation between emission and soil temperature was found with the temperature at 35 cm depth. However a spatial variation is very clear although within short distances as seen in Figure 9. Chamber 6 has emissions around 50 µmol CH_4/m2/hr, whereas emissions from chamber 1 are only around 30 µmol CH_4/m2/hr. The variation could be explained by differences in soil structure, amount of organic matter or vegetation types at the two hummocks.

A 24 hour measurement was also conducted at three chambers at DB. Unfortunately one of the chambers had a leakage and the measurements were therefore removed from the analysis. The soil temperature at 5 cm varied between 5.7 °C and 8.0 °C over the period with the maximum at 2pm. In Figure 10 greater variations within measurements throughout the 24 hours is seen compared to the variations observed at HF.

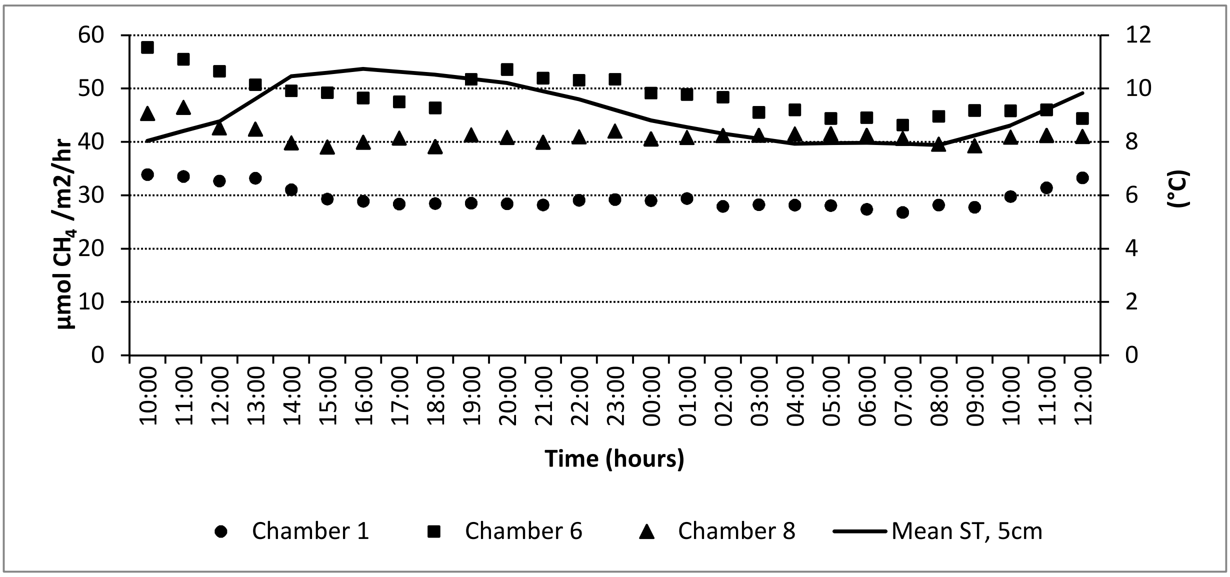

Figure 9. Results of 24 hr measurement at HF. Fluxes are given in µmol CH_4/m2/h.

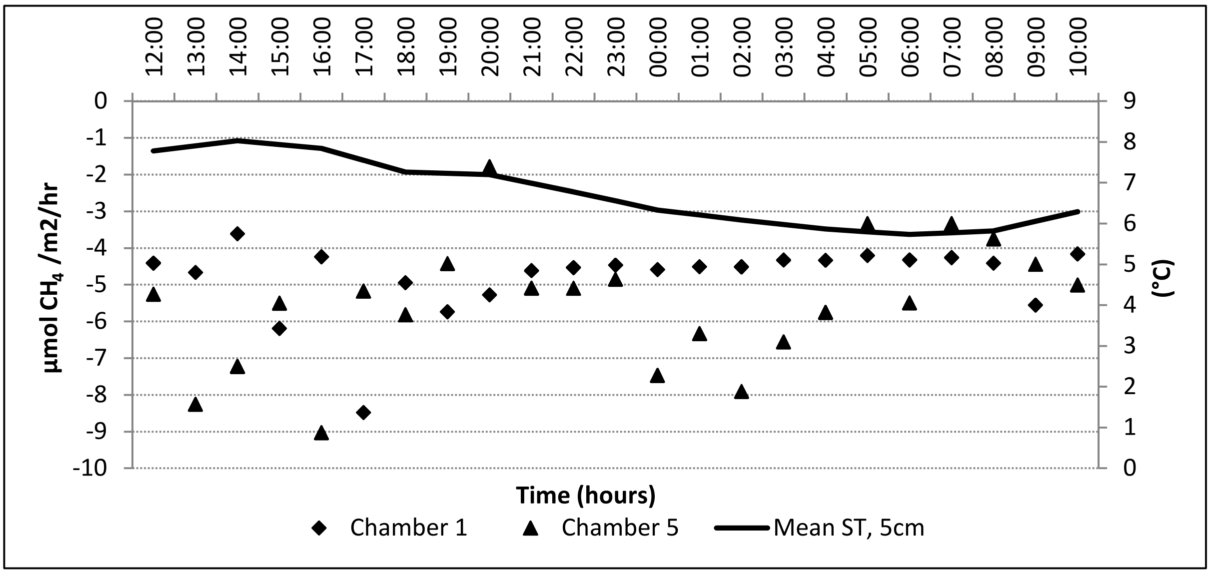

Figure 10. Results of 24 hr measurement at DB. Fluxes are given µmol CH_4/m2/hr.

Maybe spatial variations such as amount of organic material, vegetation cover or the microbial composition had an influence on these variations. Here there was also no correlation found with the fluxes and diurnal temperature variations.

CH_4 emissions from Flakkerhuk compared to other studies

The wet sites at Flakkerhuk had an average CH_4 emission of 4 mg/m^2/d and the dry sites had an average CH_4 uptake of 1.3 mg/m^2/d. The average emissions from the wet sites are very small summer emissions compared to other regions. A high Arctic site in NE Greenland measured summer emissions of almost 120 mg/m^2/d according to Friborg et al., 2000. From an Alaskan wet meadow summer emissions between 100-700 mg/m^2/d were recorded according to Callaghan et al., 2004. In Siberia midsummer fluxes were found to be around almost 18 and 30 mg/m^2/d (Jahn et al., 2010). The lower emissions from the wet sites at Flakkerhuk could be due to the specific location of the sites. All sites were placed at the youngest terrace resulting in less developed soil systems. This was indicated by the relative thin layer of peat. Secondly the area was very much dominated by stones, large pebbles and aeolian sediment which have low organic content hence explaining the lower CH_4 production compared to other areas. However the consumption rates at the dry sites at Flakkerhuk were more or less the same compared to other findings. In Alaska the consumption rates on well-drained upland sites had uptake rates ranging from 0.10 to 0.95 $mg/m^2/d$ (Billings et al., 2000). And comparing the uptake at Flakkerhuk with a study conducted in northern Europe by Dobbie et al. (1996), the rates are also quite similar. Although Flakkerhuk is categorized as an arctic environment, summer consumption rates are more or less the same as environments located at lower altitudes. The climatic and spatial differences of course restrict direct comparisons of the emissions measured at Flakkerhuk with other areas.

Conclusion

Measuring CH_4 emissions with a mobile LGR-DLT100 over different vegetation specific sites at Flakkerhuk, Disko (Greenland) showed a clear trend with wet sites emitting CH_4 to the atmosphere and dry sites consuming CH_4 from the atmosphere. No correlation was found between CH_4 fluxes and soil water content and temperature as this study is a snapshot representing the high arctic midsummer conditions. An extended measurement period would improve the understanding of net emissions of such a young system. Spatial variations in CH_4 emissions within sites were seen. One plausible explanation could be due to differences in vegetation cover and bacterial composition. It is important to emphasize that analysis of parameters such as pH measurement, nutrient status, amount and quality of organic material, and details on the microbial community and composition, probably would have been good to include in this study, to enable more detailed explanations on the spatial variations. Further studies of net CH_4 emissions in the Arctic are very essential for clarifying the feedback mechanism it will have on an already changing climate.

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Upscaling methane fluxes to a net balance for Flakkerhuk, Greenland

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Abstract. Within the same landscape both emission and uptake of methane (CH_4) can take place. Thus to determine the impact on the atmosphere it is necessary to quantify the net balance of CH_4. In this study results from chamber measurements of CH_4 fluxes representable of both sources and sinks at Flakkerhuk, Greenland are used for upscaling to a study area of 27.1 km^2. A Landsat 5 TM image is used for classifying the study area into wet and dry surface classes. The emission from wet areas is 9.86 g CH_4/day whereas the uptake on dry areas is 30.30 g CH_4/day. The study area at Flakkerhuk is therefore a net sink of atmospheric CH_4 of about 20 g CH_4/day. This uptake is only valid for the midsummer conditions that the used fluxes represent. However, the findings are subject to discussion as the area of wetlands seems underestimated and thereby underestimating the emission. The use of very high spatial resolution images for classification would considerably increase the reliability of the result.

Keywords: *Methane, net balance, upscaling, Landsat.*

Introduction

The potent greenhouse gas methane (CH_4) is one of the most important gasses in the Earth's atmosphere in response to greenhouse effect and the increasing concentration seen in the atmosphere is one of the drivers for global warming (Oldfield 2005). CH_4 is a natural occurring gas and about 30 % of the global emission is from natural sources with wetlands as the main contributor (Bonan 2008). As a natural gas it interacts in biogeochemical cycles and there is also a considerable CH_4 uptake by soils (Bonan 2008).
As presented in Johansen et al. (2011, this publication), within very short distances a landscape can act both as a source and a sink of CH_4 due to the land surface type given by landscape features, hydrologic regime and plant community. The small-scale chamber measurements were proven an efficient and reliable way to determine CH_4-fluxes. The advantage of the chamber measurements is that the source area is well defined and fluxes can thus be attributed to specific land surface types and plant communities. Thus to determine a landscape's effect on the atmosphere in response to CH_4 it is required to determine the net balance of emitted and uptaken CH_4. This can be achieved through upscaling by spatial extrapolation, where a simple way is to use point-scale measurements and multiply a mean flux for a certain surface type to the total area of that class within the study area. To perform the upscaling, information on how big an area/fraction each measured landscape class represents is required. A powerful tool for measuring the area fraction is classification of high resolution satellite images. Both the hydrological regime and the plant community are ways to separate the landscape classes and thereby the sources and sinks (Riutta et al. 2007). This can easily be done by remote sensing classification. The use of remote sensing is an easy way of upscaling to large areas and especially with the use of ground-truth points the classification can be very accurate.

With results of CH_4 fluxes obtained from chamber measurements at Flakkerhuk, Disko, the point-measurements will be used to upscale the net CH_4-balance for a local area of Flakkerhuk.

However, there is also a limitation to this upscaling method. The mean fluxes used are obtained from very site-specific chamber measurements. As the landscape of Flakkerhuk is developed through a series of uplifted ocean terraces there will be great differences in soil characteristics between the terraces. Therefore it causes a limitation for the usage of the measured mean flux when extrapolating inland

Method

Study area

The study area is situated at Flakkerhuk, Disko Island, Greenland (69°N, 53°W) about 50 kilometres northwest of the town of Ilulissat. The area is primarily flat with a gentle elevation increase inland often characterized as different aged lifted marine terraces. The surface is dominated by dwarf-shrub heath in the drier areas and wetlands where a permanent standing water table is seen. The area has a very patchy surface characterisation.

Satellite data and processing

A Landsat 5 TM scene (path 11/row 11) that was free of clouds was acquired from http://glovis.usgs.gov/ for July 2001. The spectral bands used are 1-5 and 7 with a resolution of 30 m. A supervised classification of the image was performed in ENVI based on visual determined training areas. The classification will divide the area into surface classes of wet and dry sites, which represent CH_4 emission and uptake respectively.

CH_4 flux measurements

CH_4 fluxes from two wetland sites and two sites characterized as dry were measured during two weeks in July 2011. A mean flux for each surface class was then determined. Vegetation characteristics and results are presented in Johansen et al. (2011, this publication) and these results will be the base of this study.

Procedure

The study area used for upscaling the net balance of CH_4 is selected based on topography, in-situ observations and plant communities similar to the vegetation cover at the sites where CH_4 fluxes was measured. Different soil structure, horizontal development, plant composition and hydrological regimes further inland due to landscape formatting processes presumably result in different CH_4 dynamics than measured. Therefore the study area selected will be the area between 2 and 50 metres above sea level with an extension along the coast line of 7 kilometres northeast and 5 kilometres southwest of the sampling site, see appendix. As the study area is classified in surface classes, the area of dry and wet classes is used for spatial extrapolation. However, the CH_4 fluxes used is a snapshot only representable for arctic midsummer conditions. Therefore, the upscaling conducted in this study will only give an estimate of the midsummer net balance of CH_4.

Results

Estimation of July CH_4 fluxes from fen and dry sites within Flakkerhuk are given in figure 1, where the individual fluxes from dwarf willow and dwarf birch is also seen.

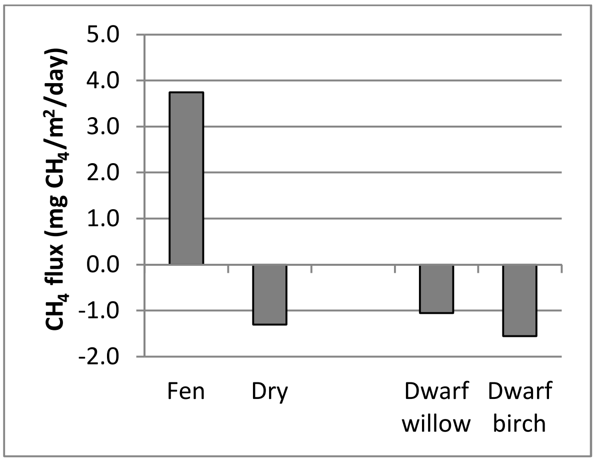

Figure 1: Measured mean fluxes by surface classes

Emissions from fen sites are estimated to be 3.74 mg $CH_4/m^2/day$ and the uptake on dry sites is 1.31 mg $CH_4/m^2/day$. Distribution of surface classes at Flakkerhuk based on the classification is listed in table 1.

Table 1: Area and fraction of classes

	m^2	%
Study area	**27147**	**100,0**
Dwarf birch	14995	55,2
Dwarf willow	8216	30,3
Fen	2634	9,7
Fell field	1130	4,2
Mud plane	172	0,6

It is seen that the dominating surface class in the study area is the dry sites with a distribution of more than 85 %. The fen area accounts for about 10 % of the area, whereas the remaining about 5 % is other surface classes where no CH_4 flux has been measured. These are therefore excluded in the further analysis. With an area of 2.6 km^2 of the study area the CH_4 emission from fens will account for 9.86 g CH_4/day. The area of dry surface classes is 23.2 km^2 and an uptake of 30.30 g CH_4/day is calculated. If the dry areas are distinguished between dwarf birch and dwarf willow the uptake is calculated to be 32.02 g CH_4/day. This means that the investigated area will act as a sink of atmospheric CH_4 of about 20 g/day. This can said to be valid for the midsummer conditions as the measured fluxes represents.

Discussion

The upscaling method determines the 27.1 km^2 study area of Flakkerhuk to be a sink of atmospheric CH_4 of about 20 g/day. This value is obtained on the basis of measurements during two weeks in July. Since the measurements are only valid for the midsummer conditions an annual CH_4 balance cannot be obtained from this upscaling. For this, longer time series is needed and especially measurements around snowmelt and freezing of the soil, as abnormally fluxes are observed in these periods (Rinne et al. 2007; Mastepanov et al. 2008).

The measure and multiply method used in this study has successfully been used in other studies (Christensen et al. 2000; Pennock et al. 2005; Schneider et al. 2009). However, this approach requires in depth knowledge of the spatial characteristics of the upscaling area. In this rather simple design of the method with 30x30 m spatial resolution data, it opens a discussion of the reliability of the result obtained. The result that the area is a CH_4 sink is also different from other studies on CH_4 net balance in arctic ecosystems, which measure such landscapes as net sources, e.g. Christensen et al. (2000).

Demarcation of upscaling area

The first problem arises whether the area chosen for upscaling is representable for the same areas as where the chamber measurements were conducted. This is important since the very site specific measurements are being the basis of surface classes for much bigger areas. As argued earlier the study area used for upscaling is therefore limited to an area, which is assumed to have similar biogeophysical properties. In particular elevation is important as the area has developed into different aged lifted marine terraces. Therefore the inland demarcation is attempted to border the second terrace as the older terraces will have a more developed soil profile and is arguably different in organic layer properties e.g. depth, C-availability and composition resulting in a different flux than measured. The demarcation is made using both in-situ knowledge and by using an elevation map. However, it is a rather subjective approach and more measurements throughout the area would have been preferable for validating the demarcation.

Classification method

High uncertainty related to the results is indisputably found in the classification, which determines the surface classes and thereby the fluxes attributed to the study area. Considerable errors can be assigned to the 30 m pixel resolution of the image that only allows a broad distinction between the surface classes. As mentioned the area has a patchy distribution of surface classes, which often means that different surface classes will be found within the same pixel in the Landsat image. The wetland area of 10 % seems much underestimated compared to what was observed during the field work. However, since wetlands mainly had a small spatial distribution they will often be included in pixels along with the dry classes, which will determine the classification of the pixel due to its apparent dominance. An improved classification could have been obtained if using very high spatial resolution images (pixel size 0.6-2.4 metres). With a more detailed classification map even small fractions of surface classes would be assigned and not just the dominating class within a 30x30 m field. Another advantage of using very high spatial resolution image would be the option of determine the training areas for the classification by ground control points (GCP). Then the exact areas where the chamber measurements were conducted would be the basis for the classification. However, this was not possible due to the spatial resolution of the Landsat image.

Conclusion

The measure and multiply method is an easy way of upscaling point measurements assigned to different surface classes to an area where the measurements are assumed to be valid for the same surface classes. Using results from chamber measurement of CH_4 mean fluxes from wet and dry sites obtained in July 2011 the net CH_4 balance for an area of the youngest marine terrace of Flakkerhuk is calculated. The 27.1 km^2 study area has an emission of 9.86 g CH_4/day and an uptake of 30.30 g CH_4/day. The area is therefore a net sink of atmospheric CH_4 of about 20 g CH_4/day during midsummer conditions. However, the findings are subject to some discussion. This be uncertainty whether the used fluxes can be ascribed to the entire study area. The most important uncertainty is the classification that determines the fraction of wet and dry areas and thereby the size of emission or uptake. Since a Landsat 5 image with a spatial resolution of 30 metres is used the classification might be too coarse to catch the patchy pattern of surface classes. Especially the wetland area and thereby the emission seems to be underestimated. Making the classification on a very high spatial resolution image and to make use of GCPs would considerably increase the reliability of the study.

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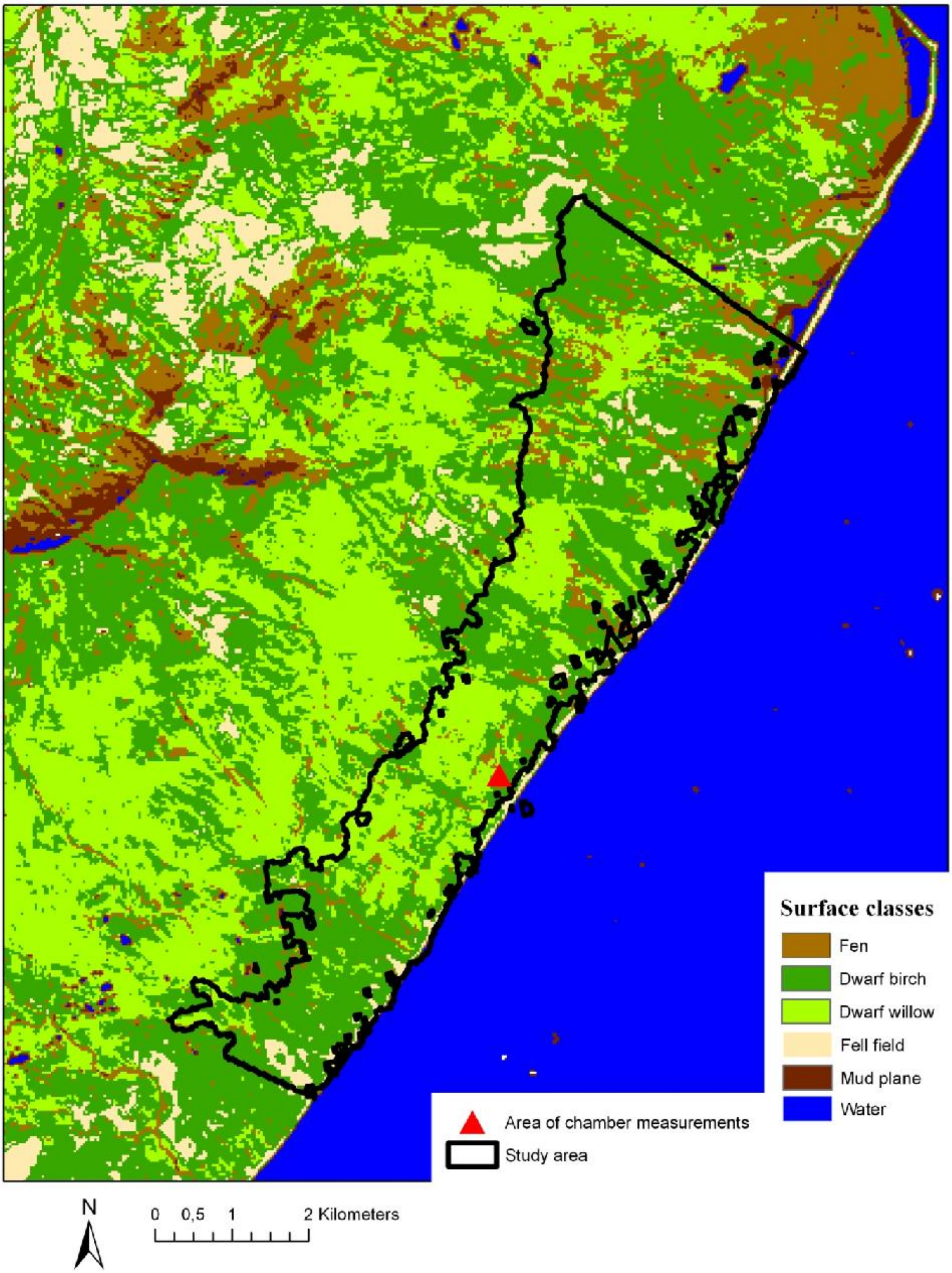
Appendix: Classification map used for upscaling



Modelling active layer thickness in arctic soils under changing moisture conditions

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Abstract. Future climate scenarios predict higher precipitation rates in the Arctic. An intensified hydrological cycle could lead to changes in soil moisture which will possibly affect the thawing rates of arctic soils. In the present study the active layer thickness is modelled for sandy soils, clayish soil and organic soil under influence of different moisture scenarios for Flakkerhuk at the east coast of Disko Island, West-Greenland. A one-layer model based on Stefans equation is applied. Thermal properties for four different soil types are analysed and used as input parameter. A multi-linear interpolation method is used to interpolate monthly temperature data for Flakkerhuk which were compared to an observed data series of a period of 11 days. The active layer thickness for dry sandy soils for July 2011 is determined by the model and results in a difference of 0.34 m compared to the observed active layer thickness. However, for higher moisture conditions the model prediction is constraint.

Keywords: *Active layer thickness modelling, soil thermal properties, multi-linear interpolation, soil moisture, Arctic*

Introduction

The predicted climate shift will lead to an intensification of the hydrological cycle caused by temperature increase. In arctic regions higher temperatures will increase the water vapour in the atmosphere resulting in higher precipitation rates on an annual basis. Recent studies predict a 20% increase of the summer rainfall over land (Rinke and Dethloff, 2008). Increasing precipitation rates imply potentially wetter soils when the soil is not frozen, which leads to a wetter active layer during summer periods (Kattsov 2007). However, if the air temperature continues to increase as seen in the last 20 year (see introduction of local climate) the evapotranspiration might increase and result in more dry soils. Trends in snow cover duration over the last 20 years (see introduction of local climate) indicate that the thawing period tends to be longer in the future due to higher temperatures. This will affect the active layer depth significantly. These tendencies will have significant impacts on the local ecosystem such as plant phenology due to changing pattern in moisture content. It can also cause thawing related greenhouse-gas emission of e.g. CO2 and CH4 (Kattsov 2007; Anisimov 2007). To which extent such scenarios will happen strongly depend on the local and regional landscape characteristics. By using model applications such feedback processes can be computed at the present state and under climate change conditions.

The present paper studies thermal properties changes of the soil under changing soil moisture conditions and different soil types. By knowing the variation of the thermal properties of the soils present in the Arctic it is possible to use this knowledge for modelling the thawing depth of the active layer for different soil types. Due to an application of an active layer model it will be investigated how changes in soil moisture affect the variation in active layer thickness.

The aim of this study is i) to find the variation of soil thermal properties of four different soil types while changing the soil moisture by stepwise saturating the soils under controlled conditions in the laboratory. The four soil types represent the dominating soil types present in the Flakkerhuk area. ii) Setting up a permafrost model on the basis of existing data and iii) incorporate the soil property from the laboratory tests into the active layer model to see changes in active layer thickness for different scenarios.

Materials & Methods

Soil thermal properties of four soil types determined in laboratory experiment

Variations of four thermal properties: thermal conductivity, specific heat capacity, diffusivity and resistivity are determined for different soil moisture contents in a laboratory experiment. The experiment was conducted on four different soil types (See Table 1). The four soil samples were collected in July and August in West Greenland. The sandy soils (S1 and S2) were collected in a profile at Flakkerhuk near the coast in a dry area. S3 was collected in a wet area (fen) further inland of Flakkerhuk in 20-30 cm depth. At its sample time it was completely water saturated as the water table in the fen was at the surface. For the fourth soil sample an organic soil was wanted but due to the young landscape in Flakkerhuk well-developed peat soils were not represented and further the soil chemical composition was out of scoop. Therefore a peat sample (S4) was collected in a fen in Kobbefjord, 30 km south-east of Nuuk. The sample contains only peat and is collected in the depth 20-30 cm. At collection time the peat was completely saturated with water.

The instrument used for measuring the thermal properties was a KD2 Pro. The KD2 Pro uses the transient line heat source method also known as heat-pulse technique. It is based on the solution of the heat conduction equation for a line heat source in a homogeneous and isotropic medium at a uniform initial temperature (Campbell et al. 1991). Two parallel needles placed in a known distance. One needle function as a heater and the other is a temperature sensor. A heat pulse gets applied to the heater and the temperature sensor records the temperature as a function of time (KD2 Pro operators manual Version 10).

The four samples were spread out on trays to air dry over a week in order to start the experiment with a water content of 0 %. During the preparation of the thermal conductivity experiment bulk density values were determined. The porosity was determined using (Table 1):

$$n = \frac{\rho_b}{\rho_s}$$

To determine the thermal response to moisture change, 1 litre of each soil type sample was poured into a beaker of 1000 ml. The procedure for each soil sample was as follows: The water was added in steps of 10. The amount of added water (g) in each step corresponded to a tenth of the porosity volume (10 measurement steps) of the 1 litre soil sample. The water was weighted prior and mixed with the soil sample in a barrel. After the soil was poured back into the 1000 ml beaker in 10 portions – each layer was compacted with an empty plastic bottle in order to keep a similar compaction and volume of the soil through the experiment. At the surface of the compacted sample three replicate measurements were taken with the KD2 Pro in order to get a mean value and standard deviation. The first measurements of the thermal properties for each soil were made in the soil sample and represent a water content of 0.

Active layer model

During warmer periods through the year the uppermost soil layer is subject to thawing. The active layer thickness (ALT) is highly time and spatial variable. The seasonal thawing is primarily defined by two climatic parameters, air temperature and snow depth as those govern annual variations of the soil temperatures. Its spatial variability on a local scale is a function of soil texture and composition, vegetation, precipitation and topography (Hollesen et al. 2011; Anisimov et al. 2002). Depending on the purpose and the model approach, more complexity to single factors can be added to create an appropriate estimate of ALT. A common approach uses a one dimensional soil-description whereas only few include time and depth dynamics of the thermal regime and soil water content within a soil column. More advanced models can be found in Riseborough et al. 2008.

In the present report the modified Stefan-solution was applied. Where the annual active layer thickness (z_t) is expressed as a function of an edaphic factor (*E*) and the temperature degree days (*TDD*) which is the amount of degrees of the daily mean air temperatures above 0 °C accumulated for all days in a year:

$$z_t = E\sqrt{TDD} \qquad \text{(eq.1)}$$

The edaphic factor is a relation of the soil physical and thermal properties such as thermal conductivity (K) bulk density (ρ_b), volumetric water content(w), and the constant latent heat of fusion (L) (333660 J/kg) to the temperature characteristics (n_t) of the area and is defined through:

$$E = \sqrt{\frac{2n_t K S}{\rho_b w L}} \qquad \text{(eq.2)}$$

A special scaling factor (S) which equals 86400 s/day is needed to get the values in the right unit.

As obtained prior to the modelling, thermal conductivity and bulk densities for each soil type are included in the modelling of ALT.

The water content is the proportion of the dry weight of the soil which is obtained by water:

$$w = \frac{Volume\ of\ added\ water}{Volume\ of\ soil} \qquad \text{(eq.3)}$$

w has the unit $m^3\ m^{-3}$.

The n_t-factor describes the relation between the soil temperature and the air temperature (see also introduction). It is the ratio of degree days at the surface (TDD_s) to the degree days of air (TDD_a) within a certain time period (Walker et al. 2011; Riseborough et al. 2008; Klene et al. 2000).

$$n_t = \frac{TDD_s}{TDD_a} \qquad \text{(eq.4)}$$

For the present model setup the observed time period of 12 days was used to compute the n_t-factor for Flakkerhuk which set to 0.8 (2010) and 0.2 (July 2011).

For modelling the annual maximum ALT the main input is air temperature which is not available for the Flakkerhuk area. These data needed to be derived prior to the modelling.

Interpolation of monthly air temperature at Flakkerhuk and TDD

Complete time series of air temperature data was available for four stations; Qeqertarsuaq (Godhavn) (Arctic Station), Ilulissat (DMI st. no. 4221), Qasigiannguit (DMI st. no. 4217) and Aasiaat (DMI st. no. 4220). The destinations and its distance can be seen in map (Appendix 1.1). The available data was on a daily basis from 01.01.2004 to 31.12.2010. As annual data is required for the modelling, it was reasonable to analyse the temperature data for each station on a monthly basis. Monthly mean, minimum- and maximum values were used to obtain a multi-linear regression analysis between these four stations. The resulting latitude (b_1) and longitude(b_2) values, R^2 and the intersection(I) values for each month were used to extrapolate the corresponding annual average, maximum and minimum temperatures for January 2004-December 2010 for Flakkerhuk. In Appendix 1.2 statistical values for the multi-linear regression analysis can be seen. An evaluation of the statistical significant cannot be given since the p-value as an indicator (<0.05) was not analysed.

To interpolate the air temperature values for Flakkerhuk the following linear equation was used:

$$T_{Flakkerhuk} = I + b_1 c_1 + b_2 c_2 + T_G - T_{avg(2004\text{-}2010)} \qquad \text{(eq. 5)}$$

where c_1 and c_2 are latitude and longitude of Flakkerhuk, respectively. The data is presented in the Appendix 1.3. The interpolated temperature data at Flakkerhuk was used to calculate the thawing degree days (*TDD*) (See paragraph

‘Local climate in Godhavn and Flakkerhuk, Disko Island’) for Flakkerhuk which is used for computing the edaphic factor. The calculation of TDD was based on Nelson and Outcalt (1987):

$$\mu T = \frac{(T_{\max}+T_{\min})}{2} \quad \text{(eq. 6)}$$

$$A = \frac{(T_{\max}-T_{\min})}{2} \quad \text{(eq. 7)}$$

$$\beta = cos^{-1}\frac{\mu T}{A} \quad \text{(eq. 8)}$$

$$T_s = \mu T + A(\frac{sin\beta}{\beta}) \quad \text{(eq. 9)}$$

$$L_s = 365(\frac{\beta}{\pi}) \quad \text{(eq. 10)}$$

$$TDD = T_s \cdot L_s \quad \text{(eq. 11)}$$

where μT is the mean air temperature, T_{max} is the monthly mean air temperature for the warmest month and T_{min} is the monthly mean air temperature for the coldest month of the year. For μT, T_{max}, T_{min} interpolated values were used. A represents the temperature amplitude, β is the frost angle where the temperature curve crosses 0 °C on the x-axis, T_s is the mean summer temperature and L_s is the corresponding length of the summer.

Model setup for four soil types

The applied Stefans solution was setup to model thawing depths for the four different soil types used in the saturation test made in the laboratory. The parameters changed in eq. 2 for calculating E includes bulk density (ρ_b), thermal conductivity (K) and water content (w) as they were all specific for each soil type.

Results and discussion

Soil physical properties

Determined soil physical properties are listed in Table 1. The rather similar samples S1 and S2 show an identical particle density with expected slight differences in bulk density and porosity values. An interpretation based on its natural occurrence is inappropriate as analysis was not based on a ring sample from the field. Only the advanced weathering or packing of the older and deeper layer S2 is may be seen in the bulk density and porosity values. Values for the clay sample (S3) and organic soil sample (S4) differ significantly from the sandy samples (S1 and S2) but its range is in the right order. Higher organic content in the peat results in higher pore space and lower bulk density.

Soil thermal properties

Figure 1 shows the mean thermal conductivity (K) measured in the laboratory for the four soil samples which were gradually saturated. All four soil types show an increase in thermal conductivity with increasing soil moisture. Thermal conductivity is the property of the soil which defines the ability of transport heat (Oke 1987). As it can be seen in Figure 1, at lower moisture contents S1 and S2 show a higher conductivity compared to S3 and S4 which are soils with the lower bulk density. In soils with a higher density there is an increased amount of contact points between the solids and it leads to larger heat flow paths (Hamdeh and Reeder 2000; Becker et al. 2003). The thermal conductivity for all soil types increases with soil moisture. For the sandy soils and clayish soil the increase is more rapid than for organic soil. Campbell et al. (1998) estimated theoretical values of K based on moisture content of different soil types. For sandy soil at the water content of 0.2 the soil

Table 1: Soil physical properties for the four soil samples.

	Sand (S1)	Coarse sand (S2)	Clay (S3)	Organic/peat (S4)
Bulk density (g cm^{-3})	1.60	1.70	1.30	0.176
Particle density (g cm^{-3})	2.60	2.58	2.10	1.29
Porosity (%)	39.6	34.1	36.3	86.3

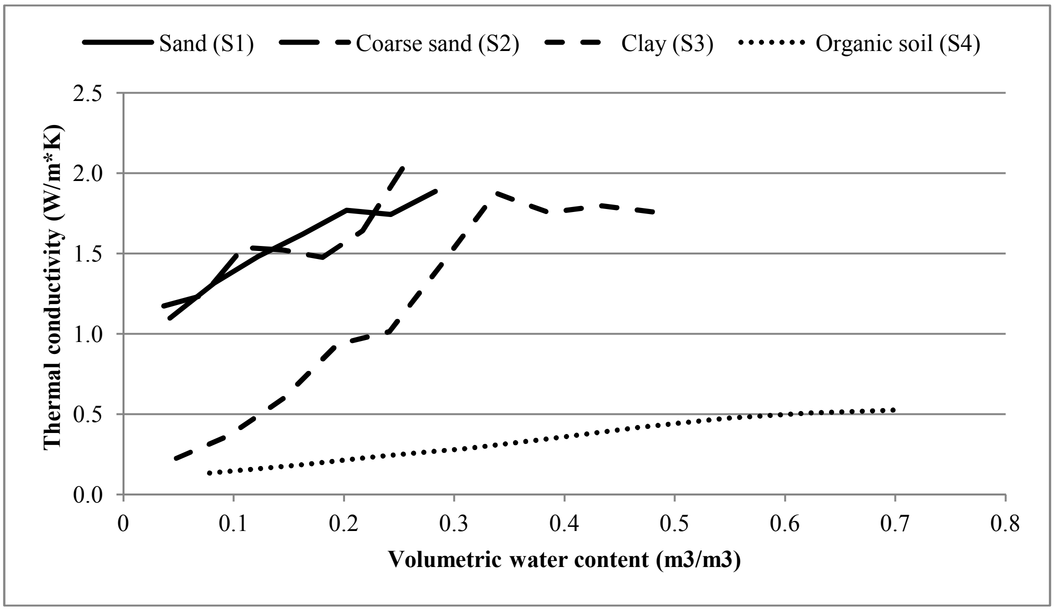


Figure 1: Thermal conductivity versus soil moisture for four soil types measured in the laboratory.

samples gave a K of 1.8 W m-1*K for S1 and 1.6 W m-1*K for S2 and , K is affected by the bulk density of soil. Campbell et al. (1998) found 1.6 W m-1*K for sand. For the similar soil moisture is K for clay 1.0 W m-1*K and Campbell et al. (1998) found 0.6 W m-1*K. This proves that our estimation of K is reasonable. K of the peat (S4) has a linear increase with changing saturation level but it never reaches the K level of clay or sandy soils but fits very well with K for clayish soils found by Campbell et al. (1998). At water content of 0.3, Campbell et al. (1998) found K for clay of 0.20 W m-1*K and K found for the soil sample is 0.25 W m-1*K.

The specific heat capacity is the amount of energy needed, to cause a temperature change in a specific volume of sub-stance (Oke 1987). Figure 2 shows the relation between specific heat capacity and soil moisture for the different soil types. The observation is not as clear to differentiate as the K-value relation. A general trend of increasing specific heat capacity with increasing saturation rates accounts for all soil types. At saturation level of 0.1, sand and clay soils do not differ significantly. However, the peat soil has a heat capacity of approximately 0.5 MJ/m3 lower at 0.1saturation level. Generally, clay has a rapid increase of heat capacity with moisture increase compared to sand. Compared to peat, clay in-creases with the same trend, but in a higher range of ~ 0.5 MJ/m3. This can be explained by the fact that with increasing moisture the clay adsorbs water and forms thick hulls around charged clay particles, which enhances its effective specific heat compared to sand soils (Hamdeh 2003). Water has a higher specific heat capacity than air and therefore more energy is needed to cause a temperature change in clay. The same effect accounts for peat. The thermal diffusivity (D) is a function of thermal conductivity (K) and specific heat capacity (C). Solids with higher thermal diffusivity react faster to a temperature change of their surroundings than solids with lower thermal diffusivity because the heat transport is faster compared to their volumetric heat capacity. Figure 3 shows the resulting thermal diffusivity (D) of the four soil types. It gives an indication that the more moist the soil becomes the more energy is required in order to change the soil temperature or thaw the permafrost. In theory, this means that if the similar microclimate is present over both a dry soil and a moist soil and both soils receive the same amount of energy from the surface, the moist soil would require a larger amount of energy than the dry soil in order to increase the soil temperature equally for both soils. Based on this theory it can then be assumed that larger thawing depths would be found in dry soils compared to moist soils if localized within the same area underlain by the same local climate regime.

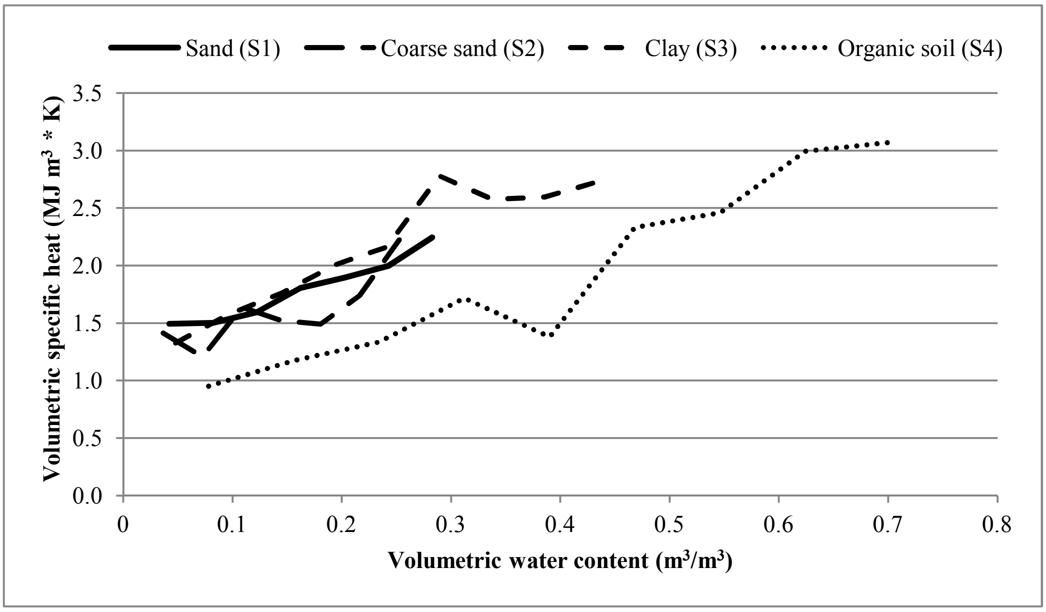


Figure 2: Volumetric specific heat versus soil moisture for four soil types measured in the laboratory.

Furthermore it can be seen in Figure 3 that sandy soils (S1 and S2) have higher thermal diffusivity than the clay and peat soil. This indicates that the coarser the soil - the higher diffusivity i.e. the ability of the soil to transport heat. This was also observed by Nakshabandi & Kohnke (1964). As well as the quartz content which has a very high conductivity compared to other soil minerals

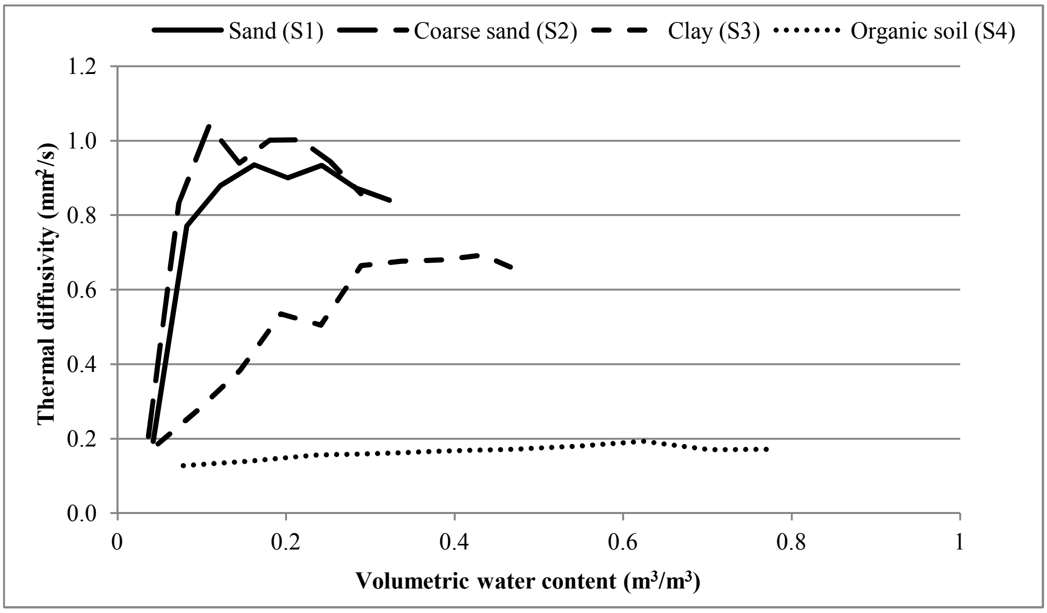


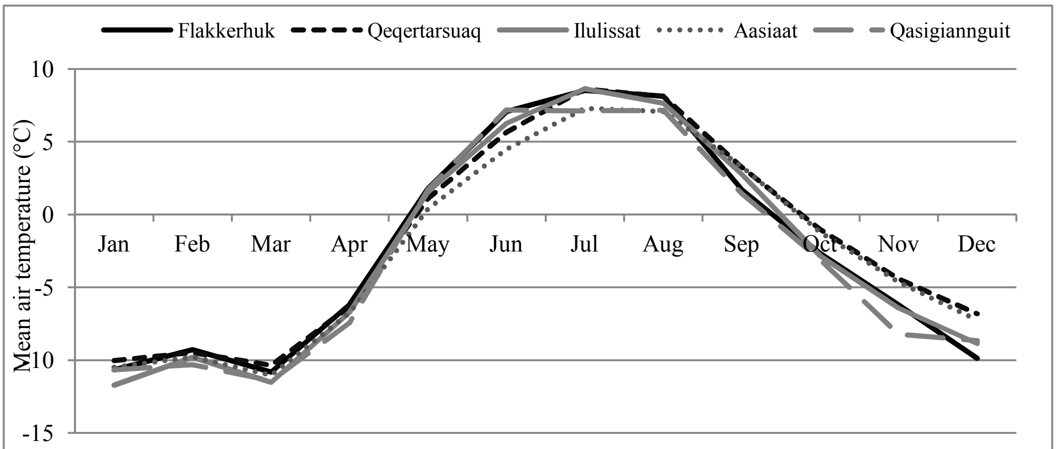
Figure 3: Thermal diffusivity found for the four soil type in the laboratory

(Campbell et al. 1998) is causing the big difference between the soil types. Hence, it is expected to see higher thawing depths in sandy soils than in soils with finer grain sizes. For the sandy soils S1 and S2 a peak can be observed at 0.10 and 0.22 water content, respectively. This is due to the relative change in K and C. Before the peak value the increase in K with moisture increase was greater than increase in heat capacity. However, after the peak the relative increase of K was lower than the constant increase in C. For the organic soil there is not a great change in thermal diffusivity with moisture. This is explained by considering the K and C curves (Figure 1 and Figure 2) where conductivity values are rather small and slowly increasing, but the heat capacity values show a rather steep increase which causes its relatively low capability of transferring heat. Meaning that the small bulk density of peat provides less contact points which lower the conductivity, but it also takes more energy to heat a certain volume of soil. Hence, it is more difficult to transport heat in such soils when it contains more moisture. In general the curves show a reliable tendency when compared to other literature (e.g. Campbell et al. 1998). D found for the organic soil and the clay soils are within the same range of D found by Campbell et al. (1998). For the sandy soils is D approx. 0.3 mm2/s higher than the values calculated by Campbell et al. (1998). This could be due to an even higher quartz content of the sand collected at Flakkerhuk. The quartz content might also be the reason for the differences between the soil samples. S1 defined the upper part of the soil column containing more organic material, whereas the lower sand layer (S2) almost only contains of quartz.

Temperature interpolation Flakkerhuk

The active layer thickness model setup requires one month with an average air temperature above 0 °C in order to calculate TDD. For normal year the latitude of 69 °N for Flakkerhuk the thawing of the soil could be initiated in May but due to the monthly mean included in the model June will be the first month contributing to the computation of TDD values. In the second half of the year thawing stops in September. Hence, the temperatures from May-September are of special interest.

From a general overview of the monthly average temperatures from 2004-2010 the difference between the stations in the first 4 months do not appear as high as compared to the last 9 months (Figure 4). Due to the difference in latitude and longitude between the four stations a dispersal of the measured temperatures can be observed from June-December. The most important parameter affecting the difference in air temperature between the four stations might be the distribution of sea ice in Disko Bay. The stations at Ilulissat and Qeqertarsuaq are located in the continental low arctic climate zone and receive cold dry air from the ice cover surface of Disko Bay. Whereas the stations in Aasiaat and Qeqertarsuaq are located near the west coast towards the Baffin Bay and the climate is influenced by low pressures moving up the west coast predominant in the summer period. This results in cloud cover in Aasiaat and Qeqertarsuaq and thereby potentially lower air temperatures compared to in Ilulissat and Qeqertarsuaq where it is most likely clear sky (Hansen 2011a). Difference in wind pattern, orientation in relation to the ocean (Disko Bay and Baffin Bay) and elevation difference between stations play a minor role. Therefore the estimated air temperature for Flakkerhuk will be



	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
T_{mean}	0.90	0.91	0.97	0.78	0.54	0.58	0.33	0.21	1.00	1.00	0.87	0.97
T_{min}	0.78	0.52	0.95	0.66	0.38	0.96	1.00	0.79	0.81	0.93	1.00	0.96
T_{max}	0.99	0.14	0.21	0.32	0.34	0.45	0.32	0.44	0.30	0.20	0.17	0.93

Figure 4: Monthly mean air temperature from 2004-2010 and R^2 values from the multilinear-interpolation. Flakkerhuk shows interpolated temperatures.

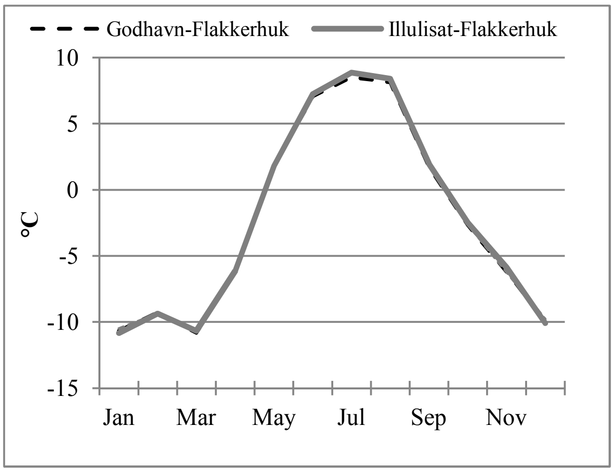


Figure5: Interpolated temperatures using different climate stations as basic input values.

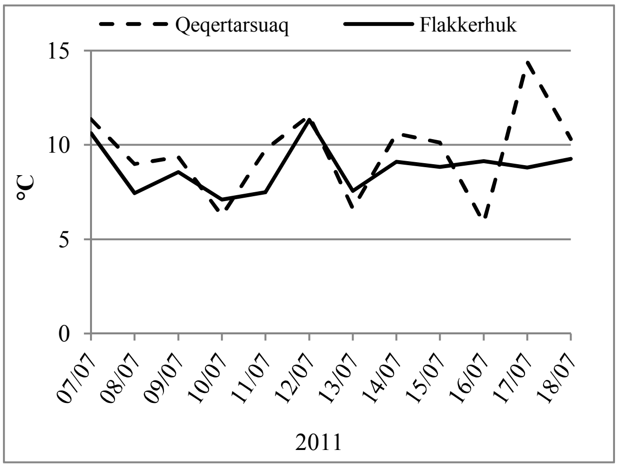


Figure 6: obtained daily mean temperatures from the observed period 07.07.2011- 18.07.2011 in Flakkerhuk and Qeqertarsuaq (Godhavn).

slightly overestimated as it is seen for June (Figure 4) where the interpolation is mainly dominated by the station in Qasigiannguit. Comparing R^2 values for the monthly average air temperature reveals that the uncertainty for the interpolation is highest in the thawing months May-August. However, the monthly minimum temperatures get rather well predicted during summer. Except for January and December the monthly maximum temperatures do not correlate well. Those trends are due to the above described differences in location within Disko Bay and lead to a general uncertainty in the interpolated data. The interpolation factors were applied to the three nearest stations (Ilulissat, Aasiaat and Qeqertarsuaq) in order to interpolate the air temperature for Flakkerhuk. The temperature curve based on data from Ilulissat and Aasiaat look rather similar, whereas for Qeqertarsuaq the estimates temperatures are lower during summer compared to the temperatures interpolated from Aasiaat and Ilulissat (Figure 5). When comparing observed air temperature from Flakkerhuk for an 11 days period (July 7-18th 2011) and observed air temperature from Qeqertarsuaq for the same period shows that it is in general colder in Flakkerhuk than in Qeqertarsuaq (Figure 6). Since the temperature data from Qeqertarsuaq were used as an inter-polation basis it should be stressed that the tem-peratures interpolated for Flakkerhuk might be overestimated in June by ~ 1.7 °C.

Active layer thickness 2010

The active layer model is tested on data for 2010 in order to get an annual run of the model. Year 2010 is chosen for the testing as it is the warmest year registered in the data series for Arctic Station. Therefore the modelled maximum thawing depths (z_{tmax}) would give an indication of the highest possible thawing depths ever reached in Flakkerhuk. The computed z_{tmax} for the four soil types sand (S1), coarse sand (S2), clay (S3) and organic soil (S4) is presented in Figure 7. In general, z_{tmax} for all soil types decreases with increasing volumetric soil water content. The dryer the soil column, the higher is the potential thawing rate of the soil. The determined z_{tmax} shows that the grain size of the soil is the main factor affecting the thawing potential at low water contents. The fine grained clay has a z_{tmax} of -1.8 m at volumetric water content of 0.1 whereas the sandy soils have a z_{tmax} of -2.55 m. The difference is explained by the lower thermal diffusivity of clay compared to sand at the water content of 0.1, see Figure 3. The sandy soils have a thermal diffusivity of 0.85-1.03 mm^2/s at water content 0.1 and whereas for clay it is 0.30

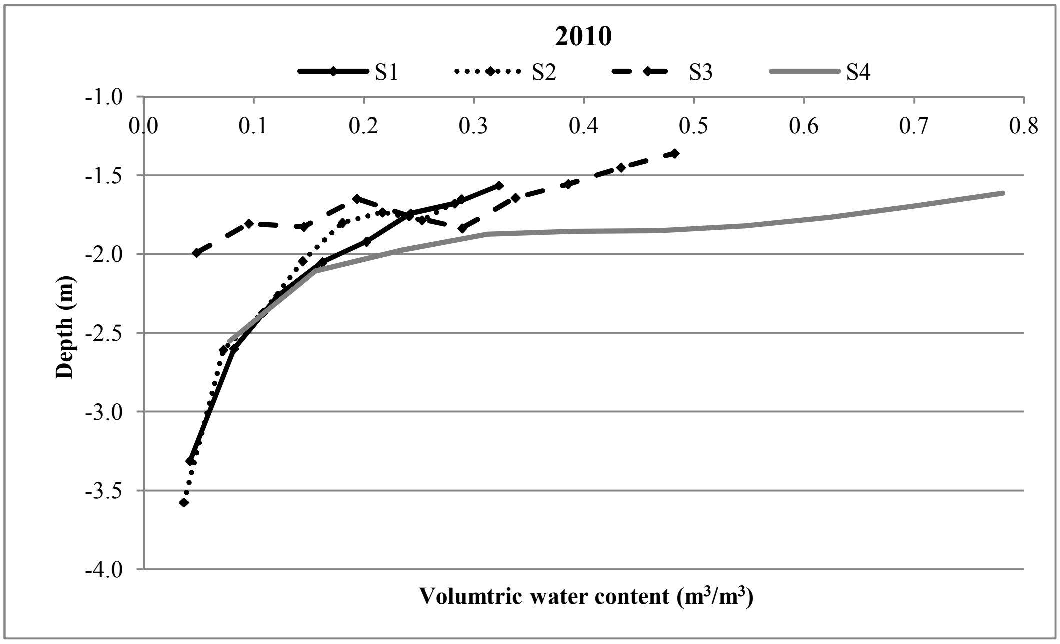


Figure 7: Modelled maximum active layer thickness for Flakkerhuk in 2010 in four soil types (S1 = sand, S2 = course sand, S3 = clay and S4= peat) with increasing water content

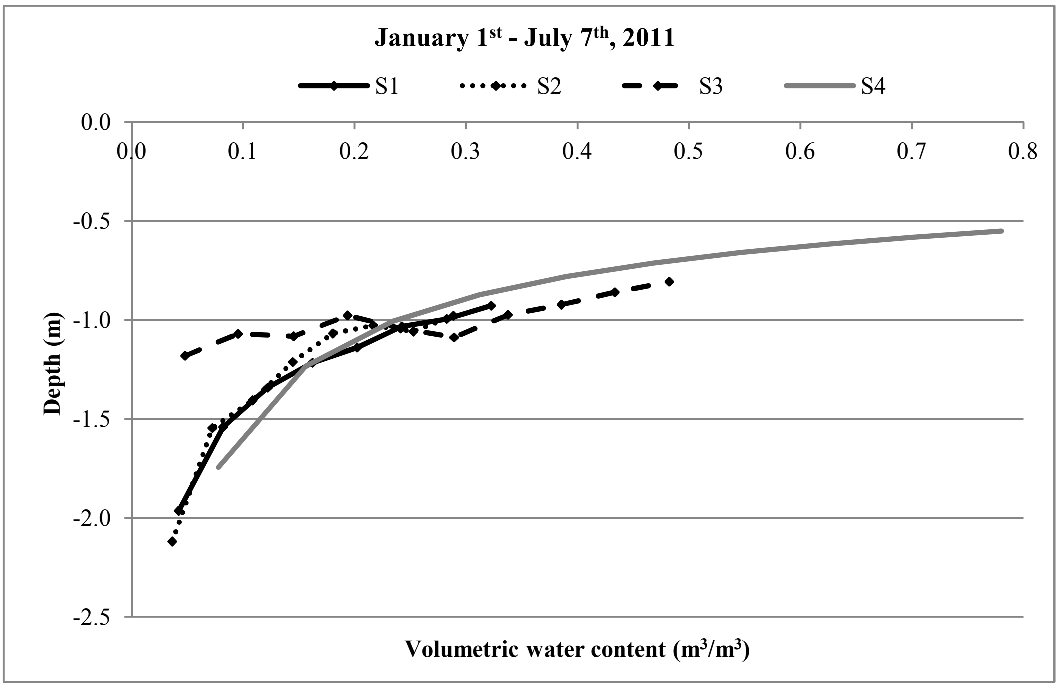


Figure 8: Figure 8 Modelled active layer thickness for Flakkerhuk at July 7th, 2011 in four soil types (S1 = sand, S2 = course sand, S3 = clay and S4= peat) with increasing saturation.

mm^2/s. This means that the energy transport in sand is easier than in clay under the same moisture conditions. Hence, a deeper thawing depth for dry sandy soil gets computed compared to dry clay soils. As the soil gets saturated gradually the difference in z_{tmax} between the soil types (S1, S2 and S3) gets minimized and it is eliminated at the soil water content of 0.3 where the sandy soils are saturated (Figure 7). For peat (S4) at a water content of 0.1 a less shallow z_{tmax} gets modelled. As the water content increases the modelled z_{tmax} decreases. When computing z_{tmax} for water contents between 0.3 and 0.8 it remains rather stabile on a depth between -1.8 and -1.6 m. This indicates that changes within this interval of soil water content, z_{tmax} does not change remarkably, but if the water content decreases to levels below 0.3 the z_{tmax} would increase exponentially. The curve of the modelled z_{tmax} for sandy soils shows a continuous exponential trend as well as steeper decrease compared to peat. This means that the sandy soil types are in general more sensitive to moisture changes and thaw up deeper compared to organic soils.

Active layer thickness July 2011

In order to evaluate the active layer model against observations of active layer thickness, the model is applied for the period January - July 2011. During the field course in Flakkerhuk active layer thickness observations were made in the beginning of July. These will be used for validating the model run for 2011.

Figure 8 shows the modelled thawing depth reached at July 7th, 2011. These results are based on 436 TDD in June 2011 and 75 TDD for seven days in July 2011. The same soil types are included as in 2010. The variation with soil moisture between the soil types is similar to the pattern seen in 2010. The thawing depths are not comparable due to the difference in timing. For 2010 is the maximum thawing depth modelled on the basis of the annual amount of TDD (1061) whereas the thawing depths for 2011 are computed until July 7th and therefore based on a smaller amount of TDD (511).

For July 7th 2011 an active layer thickness of 0.8 m was observed in a dry sandy soil (5-22 vol% soilwater) by the permafrost drilling group. The modelled active layer thickness for the same date for a sandy soil with a volumetric water content of 0.1-0.2 ranges from -1.4 to -1.14 m, respectively.

This indicates that the active layer model overestimates the values by 34-60 cm. This uncertainty is probably due to an overestimated surface temperature. Since the relation Tair/Tsurface is based on observed temperatures in July the relation is not valid throughout the whole year and not completely applicable in the colder months where snow cover is present. Hence, when applied in the model it causes too high surface temperatures. Hence, the Tair/Tsurface relation should be investigated for a longer period covering both, cold and warm seasons.

A different reason for the overestimation of thawing depths for July 2011 is explained by an overestimation of the interpolated air temperatures at Flakkerhuk, especially for the summer months. This interpolation error is due to local parameters not explained by the multi-regression analysis.

Another uncertainty is given by the calculation of TDD values which is based on monthly mean values. Daily mean temperatures would give a better estimate of this factor. For example TDD in May is underestimated as a monthly average of -0.8 indicates several days with temperatures above 0 °C. At least that is the case for the daily air temperatures in May in Qeqertarsuaq, from which 10 TDD can be calculated. In the model the starting point of thawing the ground in the model is defined by the TDD and not by the snow cover melt off. In general, the first amount of TDD would be used to melt the snow away. When the surface is snow free after several TDD the thawing of the ground is initiated. Hence, the timing of the initiation of active layer thawing could be enhanced by assigning the date of the snow free surface as the starting point for the model.

Conclusion

The study was initiated analysing soil samples and determines thermal properties under different moisture conditions. From the experiments it can be concluded that the coarser the soil type the better it conducts heat. As well as the higher moisture of the soil - the higher is its thermal conductivity and its specific heat capacity. Therefore thermal diffusivity increases abrupt and peaks for sand and clay at moisture contents of 0.1-0.2 and 0.4, respectively. The decrease in thermal diffusivity with higher moisture contents is caused by higher energy requirements in order to heat a soil column with higher water contents compared to a soil column with lower water contents.

The interpolation for the air temperature for Flakkerhuk contains several uncertainties which are mainly caused by the difference in local climate within Disko Bay. The r^2 values show that the interpolation does not describe the variation between the four stations fully on annual basis. Low correlation is especially seen during the summer months where a large variance in temperature is seen between all stations.

The modelled active layer thickness for July 2011 was compared to the active layer depth of 0.8m observed at Flakkerhuk in beginning of July 2011. The active layer model included air temperature data, soil thermal parameters and soil physical parameters. It predicted an active layer depth of 1.14 m for a dry coarse sandy soil for the date July 7th, 2011. This gives a difference between modelled and observed active layer thickness of 0.34 m. Therefore the active layer model can be considered as acceptable.

However, the applied active layer model based on the Stefan solution is only a simple 1-layer model and except for the temperature, no interaction of the soil column with its adjacent environment takes place. The soil column in the present active layer model is represented by one temperature and one soil moisture value and no further profile separation has been done. Subsurface hydrological parameters were excluded as well. Therefore it can be concluded that the Stefan solution gives a fairly good estimate when only a limited amount of climate data is available. To improve the estimates a 2-layer model approach would be favourable to apply in combination with a surface-subsurface runoff model.

Finally, based on the findings in this study it can be concluded that future changes in soil moisture content will affect the thawing rates in arctic soils. The modelled results show that the moister a soil - the thinner the active layer.

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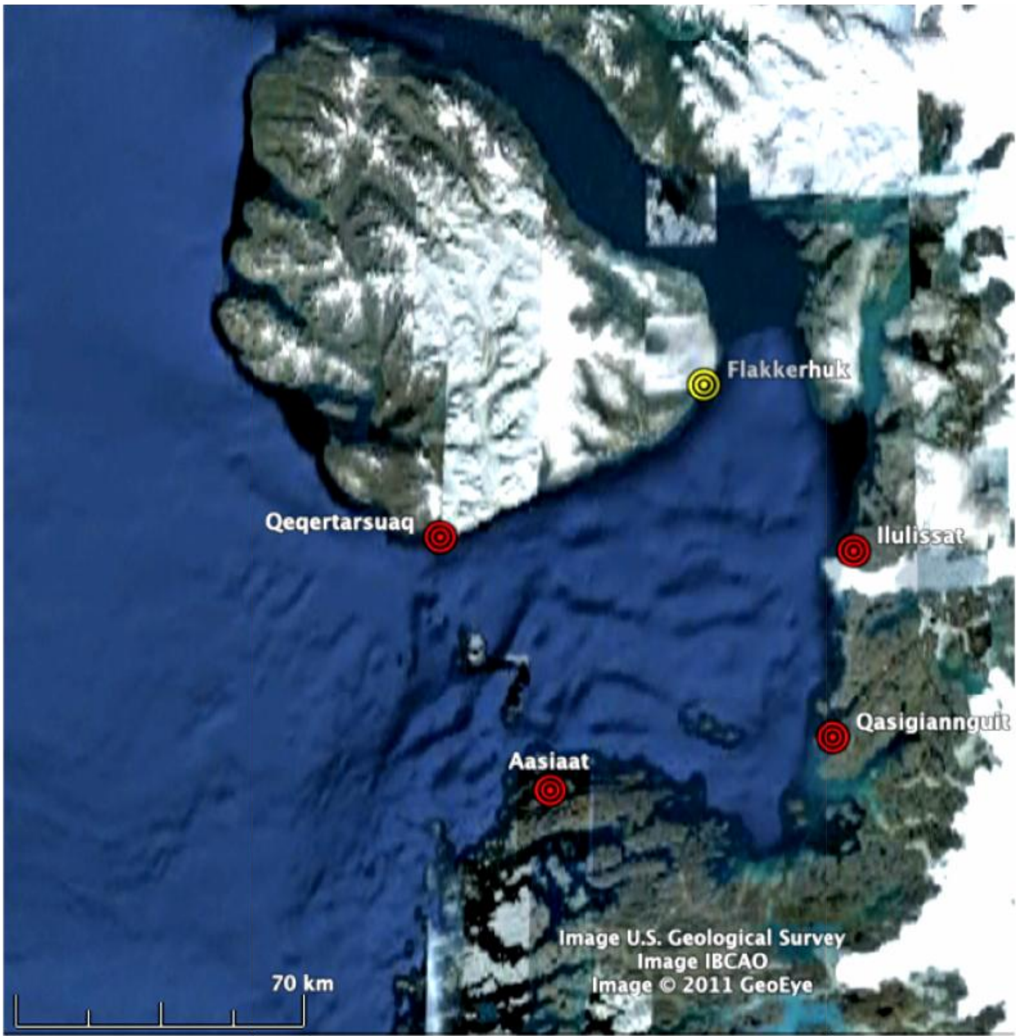
Appendix 1.1 location of climate stations used for interpolation of temperature regime in Flakkerhuk and its distances to Flakkerhuk



Distances to Flakkerhuk (km)	
Ilulisat	53,5
Qeqertarsuaq	72,7
Qasigannguit	89,4
Aasiaat	103,2

Appendix 1.2 Multi-linear regression analysis- interpolation input and R^2

T_{avg}	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Intercept	-99.56	-53.4	-34.37	-47.69	-26.26	-18.5	30.78	-25.79	36.53	7.44	64.47	40.42
LAT	0.64	0.51	-0.07	0.48	0.67	0.99	0.22	0.64	-0.77	-0.76	-1.52	-1.44
LON	-0.85	-0.16	-0.54	-0.15	0.37	0.84	0.72	0.2	-0.37	-0.83	-0.67	-0.99
R^2	0.9	0.91	0.97	0.78	0.54	0.58	0.33	0.21	1	1	0.87	0.97

Appendix 1.3 monthly average T interpolated in Flakkerhuk

Tavg	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
2004	-10.2	-10.3	-15.2	-7.3	3.0	6.3	5.8	6.7	1.1	-1.3	-5.7	-14.0
2005	-11.3	-7.9	-4.7	-5.4	0.1	7.7	8.0	7.5	0.1	-3.5	-4.4	-9.3
2006	-13.9	-8.7	-7.8	-7.7	1.9	5.5	7.6	8.2	3.5	-1.4	-8.2	-11.7
2007	-9.1	-4.3	-11.0	-7.7	0.1	7.8	10.3	9.6	2.4	-2.6	-6.2	-10.5
2008	-13.5	-16.1	-12.5	-4.7	2.5	7.9	9.7	6.4	-1.0	-5.9	-8.0	-12.3
2009	-8.2	-9.8	-16.7	-7.5	0.6	6.6	9.3	8.8	1.1	-3.5	-8.6	-6.4
2010	-8.6	-7.9	-7.8	-3.1	4.2	7.7	8.9	9.6	4.7	-0.4	-2.4	-4.9
2011	-15.6	-13.3	-10.6	-4.6	-0.5	13.1						

Climate change feedbacks of the future greenhouse gas budget

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Abstract. The active layer depth for two future scenarios for a wet fen site and a dry willow site was used to determine the potential decomposable C pool for the two scenarios respectively. The potential production of CH_4 from the sites was compared with the current CH_4 emissions from the sites. To determine the likelihood of any produced CH_4 being emitted the oxidation potential of the top soils were determined. If only the dissolved C pool is to be converted into CH_4 the top soil is able to oxidize it to CO_2. This is not the case if the solid C is fully converted into CH_4 as well. Since no differences in thawing depth were found between the wet fen site and the dry willow site the worst case future scenario is a temperature increase combined with a change in the local water regime, making the current dry soils wet.

Keywords: Methane, carbon dioxide, permafrost, methane oxidation potential, Greenland

Introduction

The increase in temperature in arctic areas caused by climate changes will have a high impact on areas with permafrost. The boundary between the active layer and the permafrost will move downwards with increasing temperatures, as the summer thawing of the soil progresses downwards. Even though increased temperatures will have a direct effect on the gas flux of greenhouse gasses (GHGs) across the soil-atmosphere interface caused by changes in the primary production and on the rates of degradation of litter in the soil, degradation of the permafrost may have an even larger effect on future GHG dynamics (Rodionow, 2006). Whether an increase in the active layer depth results in an increased methane (CH_4) flux over the soil-atmosphere boundary depends on several factors; first of all the amount of extra carbon (C) being made available for micro organisms to consume, the water content in the soil controlling the oxygen (O_2) availability combined with the status of the micro organisms in the soil and their ability to oxidize any produced CH_4 to CO_2 before it reaches the surface. The water content in the soil does not only control the O_2 content of the soil it also affects the thawing depth of the active layer, as was found in Article 5. Large increases of CH_4 emissions have been found for mires and peat soils as permafrost is thawing (Christensen et al., 2004). The aim of this study is to investigate whether this tendency also can be found in young sandy soils from the area of Flakkerhuk in Disko.

Methods

The site specifications and method description for permafrost drilling is described in Article 1 of this report. The background for the active layer depth modelling and climate change projections has been described and discussed in Article 5. The preliminary estimates of active layer depth can be seen in the Appendix. The current CH_4 surface flux in the study area have been described and discussed in Article 3. To calculate the oxidation potential for different vegetation type sites the CH_4 concentration was measured with gas chromatography, after the standard method used by Reay et al. (2005), at different incubation temperatures at three time steps. Three replicates of soil with fen vegetation and three replicates of soil with willow vegetation were collected for incubation at both 0 and 7.5 ºC. For set up of the gas chromatographer and the analysis procedure; see Christiansen & Gundersen (2011).

Results and discussion

The permafrost thawing increases with increasing temperatures (Appendix) and decreasing water content of the soil (Article 5). The wet fen site has a maximum water content of 55 % and the dry willow site has a maximum water content of 20 % (Figure 1). The water content is not uniform down through the soil but to simplify the calculations maximum values will be used to represents the sites. The differences in thawing depth between the current and the two scenarios are shown in Table 1. Scenario 1 follows the current trend of a temperature increase of 0.25 ºC where as Scenario 2 describes a 5 ºC increase (for details, see introduction of rapport).

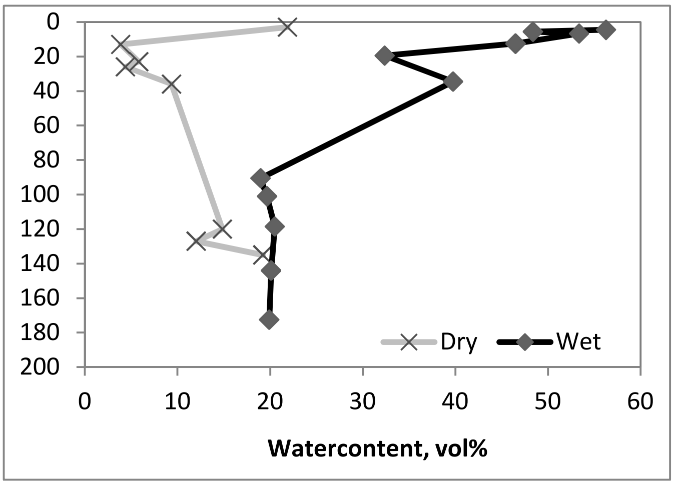

Figure 1: The depth specific water content of the dry and wet site.

The current depth of the active layer is 170 and 125 cm for the dry and the wet site respectively (Table 1). The increase in thawing depth for the two scenarios is the same i.e. 5 cm and 25 cm for Scenario 1 and 2 respectively. This deviates from the hypothesis that higher water content will decrease the thawing depth. Both soil types from the two sites are equivalent with the soil type "sand" used in the Appendix. Another soil type may cause a larger deviation in thawing depth in response to water content. As the maximum thawing depths in Table 1 are based on the model work described in Article 2, the data can only be as good as the model is. Since future scenarios cannot be validated as no real data is present, the maximum thawing depths are only estimates.

Max. thawing depth (cm)	*Wet (fen)*	*Dry (willow)*
Current	125	170
Scenario 1 (+ 0.25 ºC)	130	175
Scenario 2 (+ 5 ºC)	150	195

Table 1: Modelled maximum thawing depth for wet and dry sites at current and at the two future scenarios. Values from figures in Appendix (S1, sand).

The increased amount of dissolved and solid C made available as thawing of permafrost occurs has been calculated for the two future scenarios (Table 2). The calculations are based on the data represented in Article 1, and carried out with the same approach as was used to determine the C and N pools described herein. No DOC values are available for the dry site below 140 cm. The deepest values known for the dry site is therefore used to represent any deeper layers. This may lead to an overestimation as the dominating trend is a decline of DOC with depth. As the thawing of the active layer only occurs during the growing season, of about three months, the extra amount of C available due to permafrost melting will only be degradable in this narrow time interval.

Much more solid C will be made available upon thawing for both vegetation types than dissolved C (Table 2). The solid C pool may be much more resistant against decomposition than the readily available pool of dissolved C, making the DOC the primary factor to consider within a short time frame. The buried solid C pool is most likely slowly decomposable compounds like lignins, fats or waxes (Brady & Weil, 2008). A higher amount of dissolved C is made available in both scenarios for the wet site compared to the dry. This fits with the hypothesis that wet soils should contain more organic matter than dry soils as it is partly protected against decomposition because of a lower O_2 level, making the decomposition rates lower. C will be accumulated if the decomposition does not exceed the rate of C input. The dry soil has a higher solid content of C compared to the wet soil for both scenarios, but the values are quite close. This may indicate that the two soils previously have had more similar conditions than are seen today.

A decomposition of C will under anaerobic conditions result in formation of different compounds which may be partially oxidized (e.g. organic acids, alcohols, methane). The relationship between the different compounds will deviate based on the soil specific conditions. If oxygen is available the decomposition compound will preliminary be CO_2 (Brady & Weil, 2008). The maximum potential emissions are shown in Table 2 and calculated by assuming that all available dissolved C is converted into CH_4 in the wet anaerobic soils and to CO_2 in the dry aerobic soils. If a longer time frame is considered the solid C can also contribute to CH_4/CO_2 production as it is slowly degraded, increasing the potential emissions.

In Table 3 the actual current average flux of CH_4 can be seen. When the values of current emitted CH_4_C are compared with the future potential emission of C over a growing season it is obvious that the potential emissions exceed the current emissions thousand fold. The big difference indicates the importance of looking into other factors affecting GHG emissions.

Current average flux of CH_4	*Wet (fen)*	*Dry (willow)*
$\mu mol CH_4/m^2/h$	10.273	-2.738
$mg CH_4$_C/m^2/3months	1.0358	-0.2761

Table 3: Current average surface flux of CH_4 for a wet and dry site (Article 3).

First of all CH_4 producing organisms have to be present in the thawed permafrost, to convert DOC into CH_4. Permafrost that has been stable for thousands of years may store ancient but still viable micro organisms capable of carrying out redox reactions, e.g. increasing productivity with increasing temperatures, following the Arrhenius equation (Rivkina et al., 2004),. Whether the required micro organisms are present in the soils of this study is unknown but likely, as the present permafrost most likely has been a part of the active layer before it was buried during storm events or the like (Article 1).

Secondly, the potential emission of CH_4 from the thawed permafrost is unlikely to equal the current, as the CH_4 can be oxidized before it reaches the soil-air interface. Based on gas chromatography analysis of soils with different vegetation types incubated under different temperatures it is possible to determine the oxidation potential of the top soil. The conversion of the CH_4 concentrations measured to an oxidation potential of the top soil was done by finding the rate at time 0 for the correlation between CH_4 concentration and time.

		Scenario 1		**Scenario 2**	
		Wet (fen)	*Dry (willow)*	*Wet (fen)*	*Dry (willow)*
Additional thawing depth	*cm*	5	5	25	25
Additional DOC made available	*gDOC_C/m^2*	0.0089	0.0024	0.0621	0.0119
	gDOC_C/m^2/3months	3.761	1.003	26.124	5.014
Additional TOC made available	*gTOC_C/m^2*	0.7984	1.1877	5.7683	5.9386
	gTOC_C/m^2/3months	336.1	499.9	2428.0	2499.6

Table 2: Thawing depth and additional DOC (dissolved) and TOC (solid) made available for both vegetation types at both scenarios.

Oxidation potential for the top 3.5 cm	**Incubation at 0 ºC**		**Incubation at 7.5 ºC**	
	Wet (fen)	*Dry (willow)*	*Wet (fen)*	*Dry (willow)*
Max/Min *($molCH_4/m_2/min$)*	-0.03084/-0.00004	-0.05469/-0.00238	-0.01615/-0.00378	-0.02331/-0.00441
Mean ± 1 σ *($molCH_4/m_2/min$)*	-0.01748 ± 0.0158	-0.03345 ± 0.0275	-0.00866 ± 0.0066	-0.01300 ± 0.0095
Mean over growing season *($gCH_4\_C/m^2/3months$)*	-105.76	-202.36	-52.41	-78.66

Table 4: Oxidation potential for two top soils incubated at 0 and 7.5 ºC.

The rate is negative, as the CH_4 concentration declines over time. Because of the small data set, a linear regression was chosen with R^2 from 0.81 to 0.96. The rate was converted into $molCH_4/m^2/min$ and $gCH_4\_C/m^2/3months$. The highest oxidation potential over a growing season was found for the dry willow site incubated at 0 ºC with -202.36 $gCH_4\_C/m^2/3months$ and the lowest was found for the wet fen site incubated at 7.5 º with 52.41 $gCH_4\_C/m^2/3months$.

The variation for both sites at both incubation temperatures is quite high. This indicates that the spatial variability of soils oxidation potential is high or that there may be data errors. Since there are only three replicates for each value, it is not possible to determine if one measurement should be an outlier. But as the minimum oxidation potential for the wet site incubated at 0 ºC (0.00004 $molCH_4/m_2/min$) deviates the most from the mean value, compared to the other measurements, this may be an outlier.

The oxidation potential is about 50 % higher for both soil types incubated at 0 ºC compared to the incubation at 7.5 ºC. As all bacterial processes are depending on temperature i.e. increasing rates of decomposition with increasing temperatures, the opposite correlation were to be expected between the two incubation temperatures. In practice the samples were first measured at 7.5 ºC and afterwards at 0 ºC. Any labile available C pool should be degraded during the first incubation and not in the 0 ºC incubation, which seems to be the case here. This may be explained by a microbial lack phase in the microbial activity (Brady & Weil, 2008). If the microbial activity was slowly increasing during the 7.5 ºC incubation the full effect may first have been reached during the 0 ºC incubation, increasing the oxidation potential.

For an oxidation potential actually to decrease with increasing temperatures, the rate of CH_4 production have to have increased more. This situation is very likely in the wet fen site, where favourable conditions for methane production are present. Under normal conditions this is not expected for a dry willow site, but it may have happened in the laboratory if the water content was higher than under natural conditions (see Figure 1). The water content was in average 42.6 % for the wet site samples and 37.6 % for the dry site samples for the three replicates respectively, indicating that this may be the case.

The dry site has a higher oxidation potential than the wet site at both incubation temperatures. This is expected as the lower water content allows O_2 to diffuse into the soil and oxidize any CH_4 present. This tendency is clearest for the 0 ºC incubation, where the oxidation potential is 50 % higher. The amount of potentially oxidized C over a growing season is much higher than the potentially produced CH_4 from the DOC (Table 2). This indicates that any produced CH_4 in the future easily can be oxidized to CO_2 before being emitted and thereby not increase the current CH_4

emissions considerably. If the pool of solid C can be decomposed in the future it may produce more CH_4 than can be oxidized before being emitted. This is especially the case for Scenario 2 where the deep thawing brings high contents of C available. Whether a temperature increase in the future will lead to increased emissions of CH_4 is therefore dependent on how much of the solid C that will be decomposed and to what extent this will be converted into CH_4 or the other possible compounds earlier mentioned.

As the dry site has a much higher oxidation potential than the wet site, it is likely that any produced CH_4 deep in the soil will be fully oxidized into CO_2 before being emitted, where as only a part of the produced CH_4 in the wet site will be oxidized. Since CH_4 has a GWP 25 times higher than CO_2 (see introduction of rapport) an increased thawing of the wet site will have a larger effect on the GHG induced climate changes than a thawing of a dry site, since the increased thawing is the same. If a realistic future scenario does not only include a temperature increase but also a change in the local water regime, making more soils wet, it will result in increased CH_4 emissions. If the temperature increase is followed by a drying of the soils, the oxidation potential of CH_4 will increase and thereby decrease the GHG induced climate changes since the less severe GHG CO_2 will be the end outcome (Strack et al., 2004).

For this prediction to be true, the current oxidation potentials of the investigated soil types and all other parameters should remain the same. This may not be the case, as increasing temperatures will improve the conditions for primary production in the area, resulting in a change in vegetation type over time. Increased heterotrophic respiration during a longer growing season can change the C balance of the system, making the area into a C sink over a long term basis (Koven, 2011). An increased primary production will influence the evapotranspiration of the soil-plant system. The evapotranspiration will increase when this is combined with increased temperatures. As more water evaporates from the soil, O_2 can easier diffuse into the soil, increasing the CH_4 oxidation potential. Increased plant productivity will also increase the root system. Roots can act as macro pores, allowing O_2 to diffuse into the soil, thereby improving the conditions for the methane oxidising bacteria (Bohn et al., 2010).

For plants really to utilize the increased temperatures, plant substrate has to be available. Ammonium plays an important role in this. As the thawing increases with increasing temperatures, a new NH_4^+ pool is made available in the same way as DOC etc. When combining the potential thawing depths seen in Table 1 with the distribution of NH_4^+ down through the profiles (Figure 2) it is clear that only a small extra pool of NH_4^+ will become available. This indicates that changes in vegetation will not have the optimal conditions and major changes will only happen over a long time.

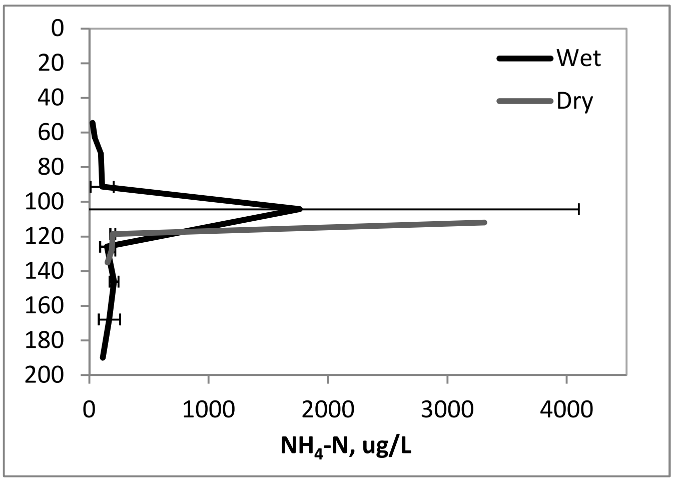

Figure 2: Depth specific content of ammonium (NH_4^+) for a wet and a dry site.

Conclusion

A future increase in temperatures will result in additional thawing of the permafrost and an increase in available C for decomposition, which can lead to an increased production of CH_4 under anaerobic conditions and CO_2 under aerobic conditions. High oxidation potentials of the top soils prevent CH_4 emissions to increase, as long as it is only the dissolved C (and not the solid C) that is decomposed in the soil. No difference in additional thawing was found between the dry and the wet site for any of the two scenarios. As CH_4 has a higher GWP than CO_2 a combination of increased temperatures and a change towards more wet soils would be the worst case scenario for increasing GHG induced climate changes. In a long term scale changes in vegetation may change the GHG dynamics completely. Increased plant productivity can enhance the O_2 supply within the soil and increase the CH_4 oxidation potential. This suggest a that CO_2 will be the primary GHG to consider under these conditions.

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Appendix

Preliminary estimates of active layer depths current at 2010 and for Scenario 1 and 2 with a temperature increase of 0.25 and 5 ºC respectively.

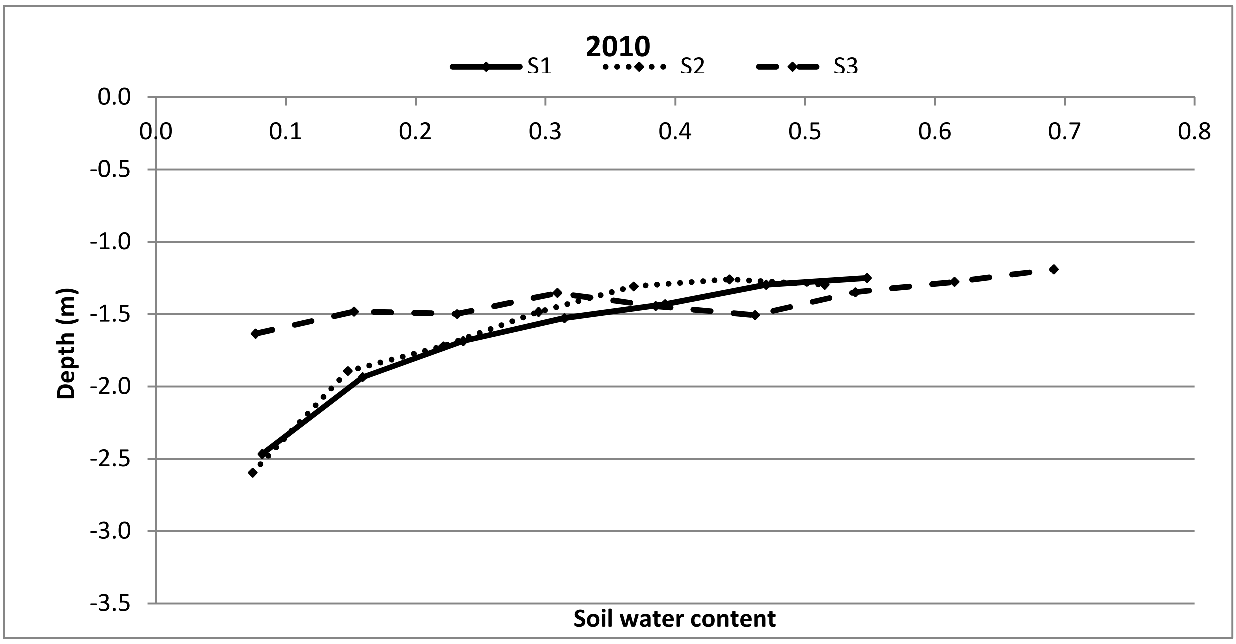

Figure 1: Modelled maximum active layer thickness in three soil types (S1 = Sand, S2 = Course sand and S3 = clay) based on mean air temperature for Flakkerhuk modelled for 2010.

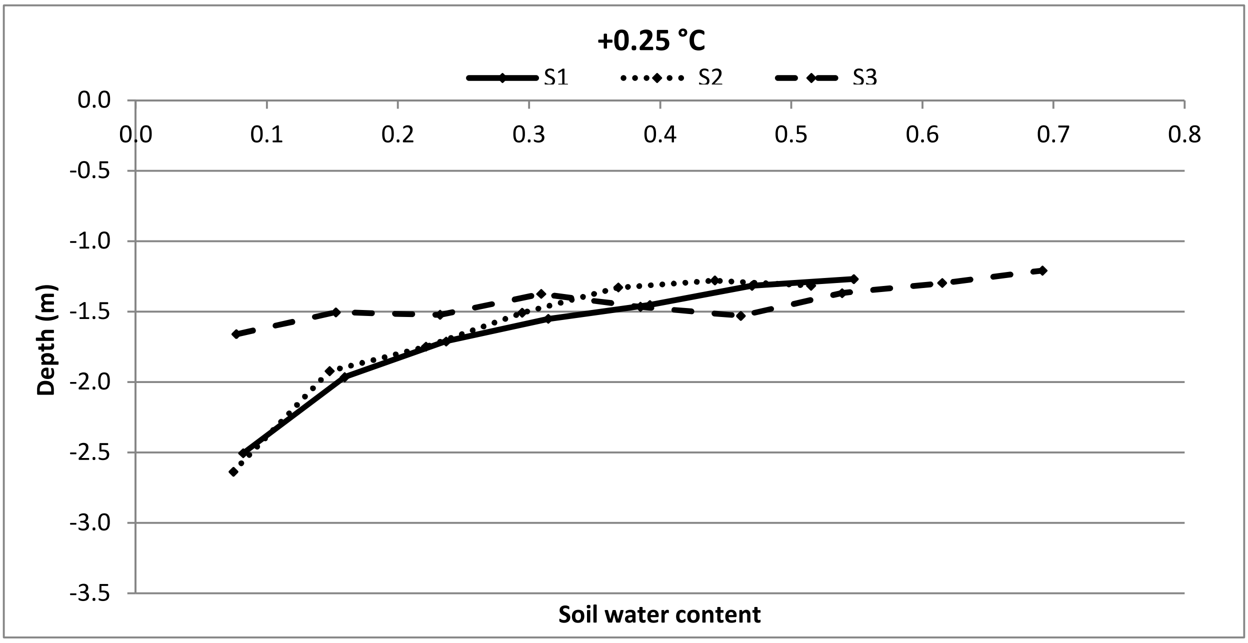

Figure 2: Scenario 1. Modelled maximum active layer thickness in three soil types (S1 = Sand, S2 = Course sand and S3 = clay) based on mean air temperature for Flakkerhuk modelled for 2010 + 0.25 °C.

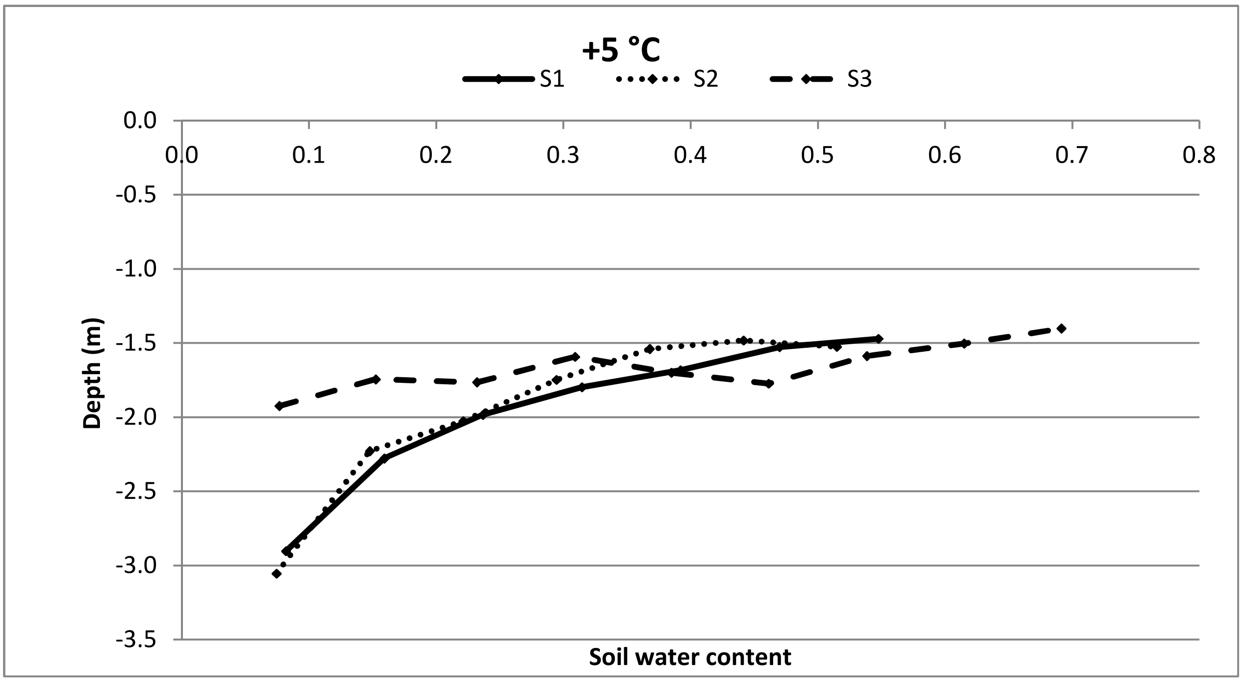

Figure 3: Scenario 2. Modelled maximum active layer thickness in three soil types (S1 = Sand, S2 = Course sand and S3 = clay) based on mean air temperature for Flakkerhuk modelled for 2010 + 5 °C.