Exchange of Radiatively Active Trace Gases in Tundra Environments, with Particular Attention to Methane

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Declaration

I hereby declare that this thesis entitled *Exchange of Radiatively Active Trace Gases in Tundra Environments, with Particular Attention to Methane* is my own original work and that all other works referred to have been acknowledged. This thesis is not substantially the same as any that I have submitted for a degree or diploma or other qualification to any other university. No part of the thesis has already been or is being concurrently submitted for any such degree, diploma or qualification. I further declare that this thesis does not exceed the prescribed length allowed me by the Faculty of Earth Sciences and Geography.

Summary and structure

This thesis is concerned with trace gas flux in tundra environments, the main subject of study being methane. Methane emission from tundra soils has in recent years attracted increasing attention due to the possible associated feed-back effect on man-made climate change. The presented data are primarily produced during field work in Northern Alaska but it also includes work in Northern Sweden and laboratory studies in Copenhagen. Model experiments carried out on basis of the gathered data was carried out in cooperation with the Hadley Centre at the Meteorological Office in England.

Chapter 1 describes the area of research and the general background for embarking on the project. It concludes by defining a number of questions which the research presented in the thesis will attempt to answer.

The first two chapters in the main body of the thesis (Chapter 2 and 3) form an introduction to tundra ecology with emphasis on aspects which are directly related to controls on soil emission of trace gases. It is explained how trace gas balances of tundra soils primarily are climatically controlled. These dependencies form main subjects of study in the thesis.

Chapters 4 and 5 contain an analysis of methane emission from tundra environments primarily based on field work in Alaska and Sweden but also involving laboratory studies of soil cores from a boreal bog. The bulk of the data presented are flux measurements produced using a static chamber technique. Methane and carbon dioxide were also analysed for their isotopic signatures. The scale of methane emission and factors controlling the flux at temporal and spatial scales are investigated and discussed in relation to the information available in the literature.

After identifying the controlling factors most useful as tools for predicting methane emission the thesis moves on to describe an attempt to model seasonal variations in flux at the main tundra site investigated (Chapter 6). This model forms part of the Meteorological Office climate prediction programme. The model is used in a number of

climate change experiments in order to assess the possible feed-back effect from tundra methane emission following different climate change scenarios.

Finally a simultaneous multigas analysis of carbon dioxide, methane, nitrous oxide and carbon monoxide flux in a tundra environment is described (Chapter 7). This forms basis for a discussion of the potential for the tundra to play a significant role in the atmospheric budgets of these gases.

In a concluding chapter the questions defined in Chapter 1 are answered based on the work presented in the six preceding chapters.

Acknowledgements

To name all the people who have helped me in the course of my thesis work would be impossible. I am grateful to all I have worked with during the past four years but I would like to mention a few people in particular.

First of all I want to express how grateful I am to my girlfriend (and coming wife) Ann-Britt Broström. The practical help she offered both in the field and with graphics for manuscripts and thesis has been invaluable. However, I am particularly thankful for her patience with me during my many departures from our home base.

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At the Meteorological Office I would like to thank Dr. David Bennetts and Dr. Howard Cattle for their support and advice on the part of the work associated with the Meteorological Office Climate Change Research Programme. I am most grateful to Dr. Chris Warner and Dr. Peter Cox whom both have shown great patience with my ignorance on the inner workings of models.

Professor Peter Williams was a perfect teacher and incredibly helpful in connection with my stay at the Geotechnical Sciences Laboratories, Carleton University, Ottawa, in 1991.

I also want to express my thanks to Dr. Søren Christensen, Dr. Per Ambus, and Anders Primé at the Department of Population Biology, University of Copenhagen. Early in the project they introduced their techniques to me and later they let me use their equipment for my own work.

Special mention should be given to Professor W.S. Reeburgh and Dr. S.C. Whalen for their help and support in connection with my field work in Alaska. Without that, this thesis would probably have taken a completely different form, if any at all. Also in Alaska I would like to thank the people I met in the wonderful community at Toolik Lake Scientific Research Station. Sarah Hobbie deserves special mention for the help she gave me in the field and Jim Laundre, and the LTER programme, for providing the weather data used in the modelling work. Special friends in Fairbanks, the Badger family (Mark, Marcia, Ashley, Birch and Claire) and Valerie Barber should not go without mentioning either.

I am grateful to Dr. Ingeborg Levin and Marcus Thom at the Institut für Umweltphysik in Heidelberg for our collaboration on the isotope work. Professor Sven Jonasson made possible the short field season I had in Abisko in 1993.

In connection with the final stages of preparing the thesis, I owe a special thank to the following for commenting on parts or the whole manuscript: Professor B. Christensen, Professor S. Jonasson, Dr. I. Levin, Professor W.S Reeburgh, and Dr. S.C. Whalen.

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Preface

My Ph.D programme has involved a lot of travelling and collaboration with research groups elsewhere than Cambridge. Consequently, this thesis contain some results from joint work with others. As stated in the declaration all data presented which are not entirely my own are duly referenced in the text. However, for the purpose of making my own contribution absolutely clear the following should be mentioned.

In the modelling work (Chapter 6) collaborators Dr. Chris Warner (Scott Polar Research Institute) and Dr. Peter Cox (Hadley Centre, Meteorological Office) carried out most of the thermodynamical and hydrological improvements of the Meteorological Office Land Surface Scheme. Only few direct results of this work are presented in this thesis but it forms basis for my development of the methane model and should therefore be mentioned here. Dr. Chris Warner and Dr. Peter Cox also carried out the actual programming of the methane parametrisation I produced.

The isotope data (section 5.3) were produced in collaboration with Dr. Ingeborg Levin and Marcus Thom at the Institut für Umweltphysik, University of Heidelberg, Germany. Marcus Thom and I did the sampling in Alaska together while Marcus Thom carried out most of the analyses in Heidelberg as described in Appendix 1. Interpretation of the data was a collaborative effort.

Part of the work presented in the thesis has already been under peer review and is published, submitted, or in the process of being so. The following is my list of publications and manuscripts connected to the thesis work.

Christensen, T. 1991. Arctic and sub-Arctic soil emissions: possible implications for global climate change. *Polar Record* 27(162):205-10.

Christensen, T. 1992. Report from the First Annual Conference of the National Institute for Global Environmental Change. *Polar Record* 28(165):162-165.

Christensen, T. 1993. Methane emission from Arctic tundra. *Biogeochemistry* 21(2):117-139.

Christensen, T. 1993. Seasonal emission of methane from the active layer of organic tundra soils - scale and controlling factors. *In*: Gilichinsky, D.A. (*ed.*) *Proceedings of the Joint Russian-American Seminar on Cryopedology and Global Change. November 1992, Pushchino*. pp:325-341. Pushchino, Russian Academy of Sciences.

Christensen, T. & Cox, P. 1993. Modelling response of methane emission from Arctic tundra to climatic change: an overview of relevant controlling factors. *Submitted to*: Oechel, W.C. & Holten, J. (eds.) *Proceedings of the Global Change and Arctic Terrestrial Ecosystems Conference, August 1993, Oppdal, Norway*. Trondheim, Norwegian Institute for Nature Research.

Christensen, T. & Cox, P. 1994. Response of methane emission from Arctic tundra to climatic change: results from a model simulation. Manuscript in process of being submitted to *Journal of Geophysical Research*.

Christensen, T. 1994. A multigas analysis of trace gas exchange between arctic tundra and the atmoshere. Manuscript in the process of being submitted to *Arctic*.

Chapter 1

Introduction

1.1. Field of research

This thesis is concerned with exchange of trace gases between arctic tundra environments and the atmosphere. The tundras of particular interest in the present work are those where considerable amounts of carbon are stored as organic matter in the ground providing substrate for substantial carbon cycling. Those areas are generally known as limited atmospheric carbon dioxide sinks and significant methane sources. This is due to the prevailing moisture regimes favouring anaerobic decomposition. Their role in the atmospheric budgets of nitrous oxide and carbon monoxide are poorly known but since tundra ecosystems generally are highly nutrient limited nitrous oxide fluxes are expected to be small.

The processes controlling exchange of the mentioned gases (CO_2, CH_4, N_2O and CO) in general, and methane in particular, between atmosphere and tundra ecosystems are main issues in this thesis. In the following sections an introduction will be given to relevant aspects related to these gases. But first a short discussion about the research area.

The study investigates geochemical, physical and climatic controls on the functioning of ecosystems and their feedback effects on the physical environment. This area of research have been named "biogeochemistry". Broadly speaking the term arises from the fact that it has become apparent there are few chemical reactions on the surface of the Earth not affected by biota. Living systems exert a major control on the composition of oceans and atmosphere. Thus, a study of the geochemistry of the surface of the Earth is the study of *bio*geochemistry (Schlesinger, 1991). Although a number of established international journals and textbooks have acknowledged the term as main focus of study it still seems to lack wider acceptance in the scientific community. It is, however, the best available broad name covering the past decades

increase in studies, within very different traditional disciplines, concerned with research covered by the term biogeochemistry. This increase has followed on the realisation that man-made changes in the atmospheric composition of trace gases could have significant direct and indirect effects on natural ecosystem functioning and, in turn, these systems might provide important feedback effects upon climate. Recent studies of these interactions are rooted in fields of research like e.g. microbiology, physiological ecology and atmospheric chemistry, fields which in their traditional definitions bears very little common ground. However, if concerned with trace gases, they all have biogeochemical implications and findings in one area could be of great importance for progress in another. It is therefore crucial that work in traditional disciplines adopt an interdisciplinary approach when concerned with trace gases in relation to issues commonly referred to as "global change". Consequently, with the risk of being accused of neglecting and "skimming" traditional scientific disciplines and in order to promote the interdisciplinary approach, I would like to see this thesis as a contribution to biogeochemistry rather than as concerned with various aspects of microbiology, geochemistry, plant physiology, soil science, etc..

The following three sections will deal with basic and historical information in relation to tundra and trace gases and the chapter will then move on to identify the questions which form main subjects of study in this thesis.

1.2. Tundra

The word "tundra" originates from the Finnish word *tunturi* which means "completely treeless heights"(Chernov, 1985). This is how the word is applied broadly all over the world to areas of higher altitude than the treeline. However, the geographically largest, and in the context of this thesis most significant tundra areas are those north of the latitudinal tree line in Eurasia and North America. These tundra areas are mainly situated in lowlands.

As is the case with the word "Arctic", "tundra" can be defined in almost as many ways as there are academic subjects concerned with it. Since this thesis is

interdisciplinary covering linkages between physical, biological and geographical aspects of the tundra an overview of various characteristics of the tundra in relation to the use of the term in this thesis will be given in Chapter 2 and 3.

1.3. Trace gases and climate

The radiative effect of trace gases CO_2, CH_4, and N_2O upon global climate is now well characterised (IPCC, 1990; IPCC, 1992). The effect is due to the trace gases reflecting and trapping infrared radiation coming back off the Earth' surface (Figure 1.1). Since the industrial revolution man-made emissions have caused an increase in the atmospheric concentrations of various greenhouse gases. The most important of the naturally occuring gases are CO_2, CH_4, and N_2O, the two former both on a global scale and as far as the possible implications from high latitude soils are concerned. Table 1.1 shows the actual changes, atmospheric lifetimes, and the relative climatic effects of these gases. CO does not have direct radiative effects. However, CO is very important for many chemical reactions in the atmosphere which influences the concentration of radiative active gases such as CH_4.

Table 1.1 indicates that the atmospheric concentrations of the gases do not reflect their relative importance as greenhouse gases. The global warming potential (GWP) of the emissions of a greenhouse gas is the time-integrated commitment to climate forcing from the release of 1 kg of the gas, expressed relative to that from 1 kg of carbon dioxide. The GWP of a given gas is dependent upon current estimates for atmospheric lifetime, present and future increase rates of the gas' atmospheric concentration and possible indirect effects of the gas associated with secondary chemical reactions. This causes uncertainties in the measure since, for example, the lifetime of methane is still controversial (Vaghjiani and Ravishankara, 1991) as well as the rate of atmospheric increase recently was shown to have lowered since the late 1980s (Steele *et al.*, 1992). Also methane has significant indirect effects which would raise the GWP quoted in Table 1.1 (IPCC, 1992).

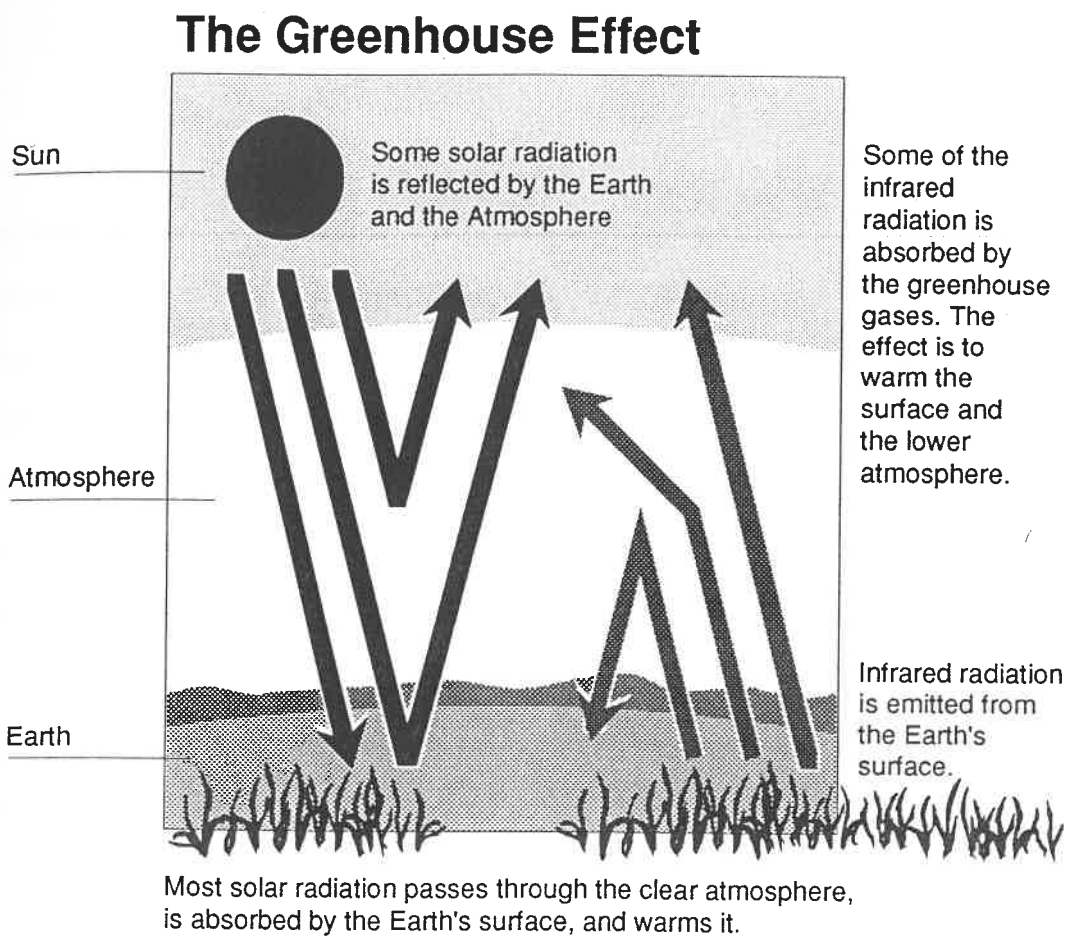


Figure 1.1. Simplified diagram showing the greenhouse effect.

	CO_2	CH_4	N_2O
Preindustrial conc.	280 ppm	0.8 ppm	288 ppb
Current conc.	353 ppm	1.72 ppm	310 ppb
Atm. lifetime (years)	120	10.5	132
Increase rate (%/yr)	0.5	0.6	0.25
GWP	1	11	270

Table 1.1. Preindustrial (1750-1800) and current (1990) atmospheric concentrations, estimated atmospheric lifetime and current rate of increase of greenhouse gases CO_2, CH_4 and N_2O. GWP is the direct global warming potential. Including indirect effects would probably raise this figure for methane.

The presence of a 'greenhouse effect' in the atmosphere determining climate was first proposed in the early part of the ninetienth century by French authors Fourier and later Pouillet (Handel and Risbey, 1992). Tyndall (1861) was the first to note that changes in atmospheric concentrations of carbon dioxide might influence climate. In an apparently little noticed paper by Hunt (1863) it was first suggested that as well as carbon dioxide also other gases including "marsh gas" (methane) could be affecting climate (Handel and Risbey, 1992) . Arrhenius (1896) provided the first quantitative discussion of the effect of carbon dioxide on climate and later made the suggestion that man-made emission of this gas could cause changes in climate (Arrhenius, 1908).

Although the emission of "marsh gas" had been well known for decades before, methane was not actually discovered in the atmosphere before the middle of this century (Migeotte, 1948). Various authors gave in the following two decades the first accounts of atmospheric methane (see Wahlen, 1993). Ehhalt (1974) made the first estimation of global tundra methane emission although no real flux measurements from tundra was used in these calculations. The first methane flux measurements in tundra-like environments were carried out as part of the tundra biome studies in the International Biological Programme (IBP). These studies included the work by Clymo and Reddaway (1971) at Moor House in Britain and Svensson (1976) who investigated a sub-arctic mire in Northern Sweden. These studies were carried out as pure "soil biology" without mentioning the application to problems related to climatic change. These issues are the background for a dramatically increasing number of studies (including the present) over the past decade. A detailed discussion of these later studies will be found in Chapters 4 and 5.

1.4. Global climate change and significance of the Arctic

Global circulation models (GCMs) are used to predict what effect the changing atmospheric concentrations of radiative trace gases will have on global climate. A situation where the atmospheric concentration of CO_2 has doubled the preindustrial

concentration (a '2 x CO_2 scenario') is normally used in the predictions. According to the most recent models this doubling time varies between 60 and 100 years from present (IPCC, 1992). There are two types of models: equilibrium and transient models. The latter are normally coupled with atmosphere-ocean interactions. The equilibrium models do not allow for any effect of gradual change in concentrations with time as the transient models do. Effect-delaying features such as the thermal inertia of oceans are also not included in the equilibrium models. However, they are more widely used and there are many more of them to compare. Transient models with coupled atmosphere-ocean are still at a relatively early stage of development. In a recent review of those that exists (IPCC, 1992) it was shown that the transient results changed little with the overall earlier conclusions (IPCC, 1990) based on equilibrium models. The most important differences occured in the southern high latitude oceans where warming was retarded due to the buffering effect of the circumpolar antarctic current allowing no warmer currents from north to penetrate far south. Warming was also lower in the northern North Atlantic where deep water is formed. However, warming over northern landmasses generally showed the same results in equilibrium and transient model experiments.

GCM models suggest an equilibrium global mean temperature rise on doubling of CO_2 of 1.5 and 4.5°C with a "best guess" at 2.5°C (IPCC, 1990). The corresponding transient results vary between 1.5 and 2.5°C (IPCC, 1992). However, this warming will not be equally distributed. More warming is expected at higher northern latitudes; lower latitudes are expected to see very small changes, if any. The Arctic figures prominently primarily because of the expected reductions in sea ice and snow cover and the effect this will have on the reflection of solar radiation from the surface (albedo, see Chapter 2).

Temperature changes in arctic terrestrial environments under 2 x CO_2 scenarios are expected to be around 4°C in winter and 2°C in summer (IPCC, 1990; IPCC, 1992). There are regional variations in and between the models but general agreement that

some of the most substantial warming on Earth will occur in central regions of northern North America and Eurasia where the bulk of global tundra areas are situated.

Precipitation changes following warming scenarios are more uncertain than the temperature predictions. However, most models predict increased precipitation throughout the year. How the combined effect of warming and increased precipitation will translate into changes in soil moisture and thaw depth is even more uncertain. However, as will be shown repeatedly in the following chapters, this is at the same time of crucial importance for possible feedback effects from tundra environments. Much attention has rightfully been given to the feedback effects from reduction in extent and thickness of sea ice and ice sheets in polar regions. Comparatively little is known about possible feedback effects from arctic terrestrial environments which is the central incentive behind this thesis.

1.5. Questions

The concept of having a largely climatically controlled trace gas balance in an environment subject to future significant climatic change was the overall incentive behind embarking on the present study. The following broad questions were defined. An attempt to answer and discuss them based on the work presented in this thesis will be given in Chapter 8.

What are the general physical and biological characteristics of tundra and how are they interlinked with climate?

How do these characteristics affect the carbon cycling in tundra ecosystems?

What are the present rates of methane emitted from true arctic tundra?

What is the isotopical signature of methane emitted from true arctic tundra?

How does this emission compare with more extensively surveyed tundra-like environments in the Sub-Arctic?

Do the characteristics of methane emission from true arctic tundra correspond with assumptions made about them in recent attempts to estimate the atmospheric methane budget?

What are the controlling factors on net methane flux from tundra soils to the atmosphere?

Can the most important of these factors be given priorities and quantified in a way that allows predictive models to reproduce seasonal variations in methane flux?

If so, then how will modelled tundra methane emission respond to climatic change scenarios predicted by GCMs and what are the major uncertainties in such predictions?

What are the general roles of tundra regions in the atmospheric budgets of other trace gases with relevance to climatic change (CO_2, N_2O, CO)?

Are there reasons to believe the dynamics of these gases in tundra environments could provide feedback effects upon climatic change?

Chapter 2

Physical and biological characteristics and extent of tundra: an overview

2.1. Physical characteristics

2.1.1. Climate and vegetation

Natural zonation of biomes like tundra and taiga, on a global scale, is primarily the result of differences in solar radiation. For example, annual totals of solar radiation measured at the ground surface in North America, decrease from about 4200 MJ m^{-2} yr^{-1} at 50°N to 3100 MJ m^{-2} yr^{-1} at 75°N (Barry *et al.*, 1981). However, not all of the total radiation (K) is available on the ground surface. On a local scale, pronounced differences in ~~total~~ short wave radiation $K(1—a)$ arise from spatial changes in surface albedo (a). Albedo ranges from 0.9 for fresh snow, to 0.1-0.25 for most vegetated or soil/rock surfaces, and can be as low as 0.05 for water bodies.

The sum of the total energy budget at the surface also includes the balance of incoming and outgoing infra-red radiation (L^*) and, thus, the net radiation (Q^*) is given by

$$Q^* = K(1—a) + L^*$$

which represents the energy available at the surface for the different heat transfer processes:

$$Q^* = Q_H + Q_{LE} + Q_G$$

where Q_H = sensible (atmospheric) heat flux, Q_{LE} = latent (evaporative and condensative) heat flux, Q_G = soil heat flux. Typical annual tundra values of Q^* are 400-800 MJ/m^2 increasing with decreasing latitude (Chernov, 1985);Table 2.1.

Observation area	Total radiation (MJ/m^2)	Net radiation (MJ/m^2)	Albedo %
Tundra	2804	670	50
Northern taiga	3307	1046	38
Southern taiga	3641	1339	35
Steppe	4981	2051	30

Table 2.1. Mean annual radiation indices from different ecosystems along a cross-latitudinal line from Dikson (73°N) to Kyzyl (51°N) in Siberia. Modified from Chernov (1985).

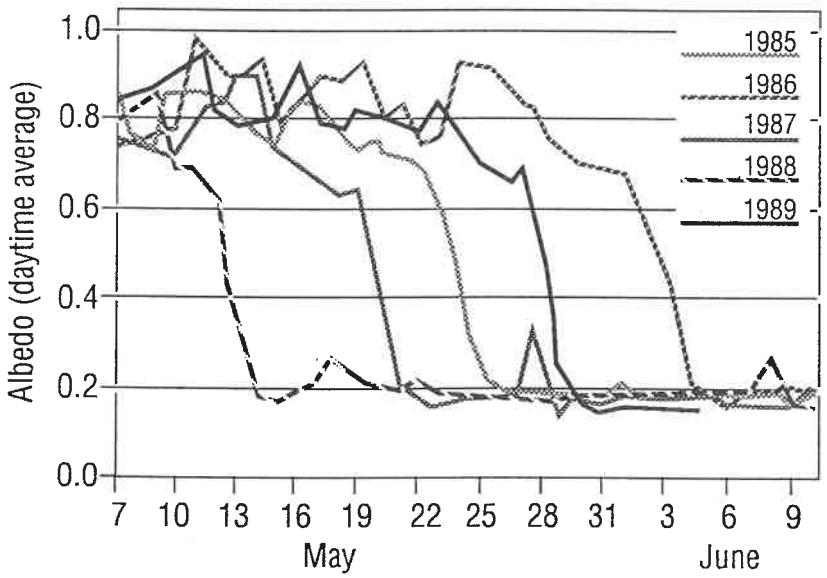


Figure 2.1. Change in surface albedo as a consequence of snow melt and variation in timing of melt for a low arctic water shed over five years (from Kane *et al.*, 1992).

The partitioning of the available energy (net radiation) at the surface shows considerable seasonal variation, mostly related to timing and distribution of snow cover. There is typically a rapid snow melt in May or early June on the arctic tundra. The snowmelt is accompanied by a sharp decrease in albedo (Figure 2.1; Barry *et al.*, 1981), which causes an increase of net radiation by an order of magnitude although the timing of thaw varies considerably from year to year (Figure 2.1). The "isolines" of net radiation agree better with vegetation zones and subzones than do latitude, because net radiation integrates the effects of total radiation and snow cover (Chernov, 1985).

Net radiation is therefore a key factor controlling life on the tundra. Temperature is closely related to net radiation and should be a good indicator of conditions for biological processes. However, it is not so much the extreme temperatures that determine the conditions in the Arctic. The coldest areas in the north are not found in the tundra but in continental areas of the taiga. For example, Verkhoyansk in Siberia - known as "the pole of cold" - is situated south of the tundra and is cold because of its strong continentality.

As mentioned above, the yearly amount of heat supply depends on net radiation at the ground surface. For living organisms, the primary role is determined by the summer temperature, which can be quite high on the tundra. Temperatures of 20 to 25 °C for several days are not uncommon, and in Siberia summer temperatures of 29°C has been measured at latitudes as high as 73°N (Chernov, 1985). However, maximum temperatures are not the most critical factors for living organisms. Instead, the number of days with temperatures within the range suitable for biological activity is more crucial. Patterns of plant and animal associations are, hence, best correlated with accumulated annual degree-days above a certain critical limit. Fairly good approximations of the boundaries of different major vegetation zones can be derived from contours of annual degree-day totals above 0 or 5°C ("growing degree-days"), although the relationship may break down where precipitation limits growth (Maxwell, 1992). A related indicator is the mean July temperature. This is often used to subdivide the Arctic because different isotherms roughly correspond with major vegetational

changes. For example, the mean July 10°C isotherm coincides largely with the forest-tundra/true tundra boundary, also quoted as the boundary between the Sub- and Low Arctic. Similarly the southern border of the High Arctic corresponds roughly with the 5-6°C mean July isotherm (Maxwell, 1992).

The soils of the tundra region are generally wet, and a large proportion of the global wetland area is situated within the tundra zone. This is, however, not due to any large amounts of precipitation. If moisture were to be defined only in terms of precipitation, large tundra areas would fall within the most precipitation-deficient of landscapes (Table 2.2; Chernov, 1985). The high soil moisture content is a result of low evaporation and drainage rates. The low rate of evaporation is, in turn, a consequence of insufficient heat, while the low drainage primarily is caused by the presence of permafrost which inhibits vertical movement of water.

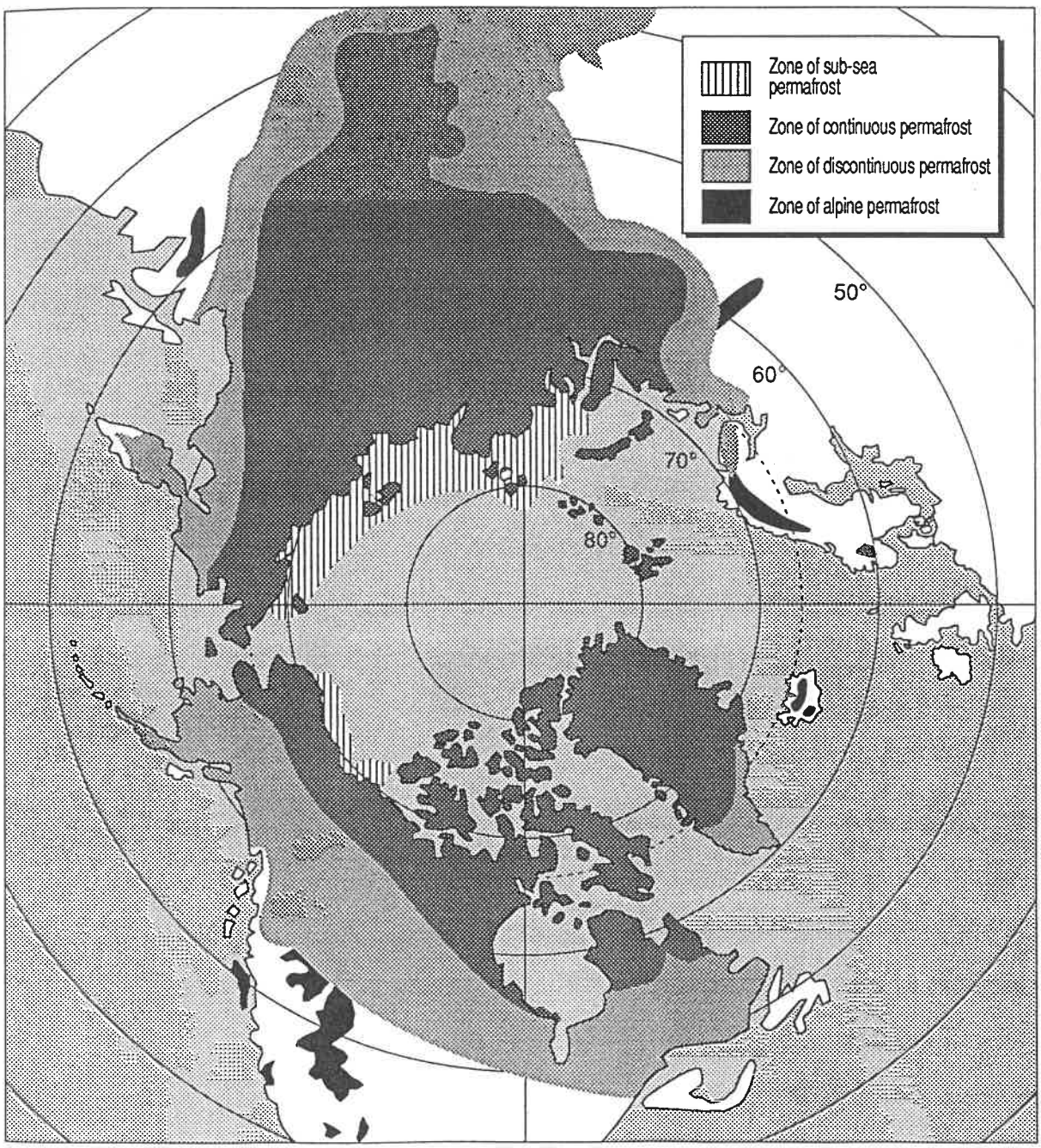
Table 2.2. Precipitation in different ecosystems (Chernov, 1985). As comparison the mean annual accumulated precipitation in Cambridge is 560 mm.

Ecosystem (or "zonal type of landscape")	Precipitation (mm/yr)
Polar desert	110
Arctic tundra	200
Typical tundra	250
Taiga	370
Mixed forest	450
Forest steppe	380
Desert	100
Tropical forest	1000

Temperature and moisture are closely interdependent in both the soil and at the soil/atmosphere interface. The higher the temperature, in general the higher are rates of evaporation. Evaporation, however, leads to cooling by the release of latent heat. Also,

the soil thermal regime is related to soil moisture and determines the depth of the seasonally thawed layer (the active layer). The latter, in turn, influences both the soil moisture regime and the soil temperature. The complexity of interactions between the active layer and the permafrost dynamics will be illustrated below.

Figure 2.2. Circumpolar distribution of permafrost.



2.1.2. Permafrost and soil climate

Permafrost. In terrestrial regions of the Arctic, a zone of permanently frozen soil, occurs below the soil surface (Figure 2.2). The presence of underlying permafrost is one feature that always distinguishes the tundra from treeless ecosystems of the temperate belt. Each summer a shallow soil horizon thaws as a consequence of a positive heat flux at the ground surface. Permafrost persists because this active layer completely refreezes in winter, removing all geothermal heat and allows the permafrost to remain in relative equilibrium with the climate. The active layer is shallow, but it plays a crucial role in tundra hydrology and is also the site of biogeochemical cycling, which is the main subject of this thesis.

Permafrost forms, per definition, where the mean annual ground surface temperature is maintained below 0°C. However, differences in microclimate, accumulated organic material, snow and the vegetation cover, influence the near-surface temperature regime, which in turn affects active layer depth, soil climate and hydrology.

Ground thermal regime. In order to describe the changing heat content of a substance, such as soil, it is necessesary to know its heat capacity. The mass heat capacity, c (J kg^{-1} K^{-1}), is the amount of heat required to change the temperature of 1 kg of the substance by 1 K. For a given amount of heat supplied, changes in temperature will be greater in a material with a low heat capacity. The mass heat capacity multiplied by the density of the substance, r, gives the volumetric heat capacity, C, measured in J m^{-3} K^{-1}. In changing its temperature by an amount dT, a unit volume of substance will experience a change in heat content of CdT. In the case of composite materials such as soils, the following weighted average value for heat capacity, C_s, must be used (Williams and Smith, 1989):

$$C_s = X_m C_m + X_o C_o + X_w C_w$$

where X is the volume fraction of soil minerals (m), organic material (o) and water (w). Table 2.3 shows how the heat capacity varies with materials and water content.

Soils freeze over a range of temperatures and unfrozen water exists in soil pores at temperatures several degrees below 0°C. The ratio of frozen to unfrozen soil moisture combined with the effect of freezing on the bulk density of the soil has significant effects on the heat capacity of soils. The rate of change of frozen content with temperature can be calculated using a simple analytical expression and used to define an 'effective' heat capacity which varies with moisture content and temperature (Chapter 6, Williams and Smith, 1989). Thus, soils with some frozen water behave as if they have an extremely high heat capacity. The presence of unfrozen water in freezing soils also has consequences for hydrology. Soil suction and hydraulic conductivity in freezing soils can be approximated by replacing total moisture content with the unfrozen moisture content in the respective standard equations (Williams and Smith, 1989). This is not an exact description but it has the correct functional behaviour and correctly predicts the enormous suctions which result from freezing the majority of the water content in a given soil. Chapter 6 describes a model which applies these representations of freezing soils.

Rouse *et al.* (1992) developed a model to trace the effect of summer warming on the energy balance (including maximum thaw depth) of tundra soils. Their model provides a useful example on which factors exert the major influence on the maximum active layer depth. They calculated thermal conductivities (k_S) and diffusivities (κ) from Q_G (see section 2.1.1), soil temperature profiles, DT_S/Dz, and heat capacity, C_S, where:

$$k_S = -Q_G/DT_S/Dz,$$

and

$$\kappa = k_S / C_S$$

The rate of downward movements of the thaw depends on the energy input at the surface (air temperature) and the thermal diffusivity (κ). The latter is a measure of the

time required for a change in surface temperature to move downward in the soil. As will be apparent, soil moisture determines the magnitude C_s and thus exerts a major influence on κ (Rouse *et al.*, 1992).

Table 2.3 Thermal properties of soils and their constituents (from Williams and Smith, 1989).

		Density (kg/m^3)	Mass heat capacity (J/kg/K)	Thermal conductivity (W/m/K)	Thermal diffusivity (x $10^{-6} m^2/s$)
Quartz		2660	800	8.80	4.14
Clay minerals		2650	900	2.92	1.22
Organic matter		1300	1920	0.25	0.10
Water (0°C)		1000	4180	0.56	0.13
Ice (0°C)		917	2100	2.24	1.16
Air		1.2	1010	0.025	20.63
Unfrozen soils:	Water content (m^3/m^3)				
Sandy soil	0.0	1600	800	0.30	0.24
(40%	0.2	1800	1180	1.80	0.85
porosity)	0.4	2000	1480	2.20	0.74
Clay soil	0.0	1600	890	0.25	0.18
(40%	0.2	1800	1250	1.18	0.53
porosity)	0.4	2000	1550	1.58	0.51
Peat soil	0.0	300	1920	0.06	0.10
(80%	0.4	700	3300	0.29	0.13
porosity)	0.8	1100	3650	0.50	0.12

This identifies two major temporally changing factors influencing the ground thermal regime and active layer depth: energy input and soil moisture, in addition to the above mentioned effect of spring snowmelt on net radiation. In winter, the snow cover forms a layer of low thermal diffusivity between air and ground, which serves to insulate the ground from the extreme temperatures of the air. Since heat exchange takes place at the snow surface rather than the ground surface, the range of annual ground surface variation is reduced and ground temperatures are higher than soils under similar climate but without snow cover, not only in winter, but also on an annual basis. Thus, where annual mean temperatures are close to 0°C, the insulation by a snow cover can impede the formation of permafrost in certain locations (Williams and Smith, 1989). Conversely, peat, which also acts as an insulator, is commonly associated with the existence of permafrost at certain locations in areas with annual mean temperatures close to 0°C. Its conductivity varies seasonally with moisture conditions (Table 2.3). When it dries out in summer, the conductivity is low and the soil beneath is shielded from the high air temperatures. Consequently, the mean annual ground temperatures tend to be low. Even if the peat remains wet during the summer, the evaporation will lead to loss of heat and, hence, act to keep the temperatures low. Apart from generally lower soil temperatures compared to mineral soils, highly organic tundra soils also exhibit smaller diurnal variations (Figure 2.3).

Surface vegetation also imposes important insulating effects. In winter snow tends to accumulate in areas with shrub vegetation which can have the warming effect mentioned above. In summer, however, vegetation might provide a similar insulating effect on permafrost as peat.

The last factor influencing the tundra ground thermal regime to be mentioned here is the presence of water bodies. Even at very high latitudes most bodies of water (except small ponds) do not freeze to the bottom in winter and this has a marked effect upon ground temperatures and the configuration of permafrost (Williams and Smith, 1989). Since the bottom temperature must be greater than 0°C the water bodies constitutes heat sources giving rise to anomalous heat flow and temperature conditions

in the ground. It is such warmer soil conditions, with the associated absence of permafrost, that causes tree and shrub growth along river banks to penetrate further north than corresponding vegetation between watercourses (see later).

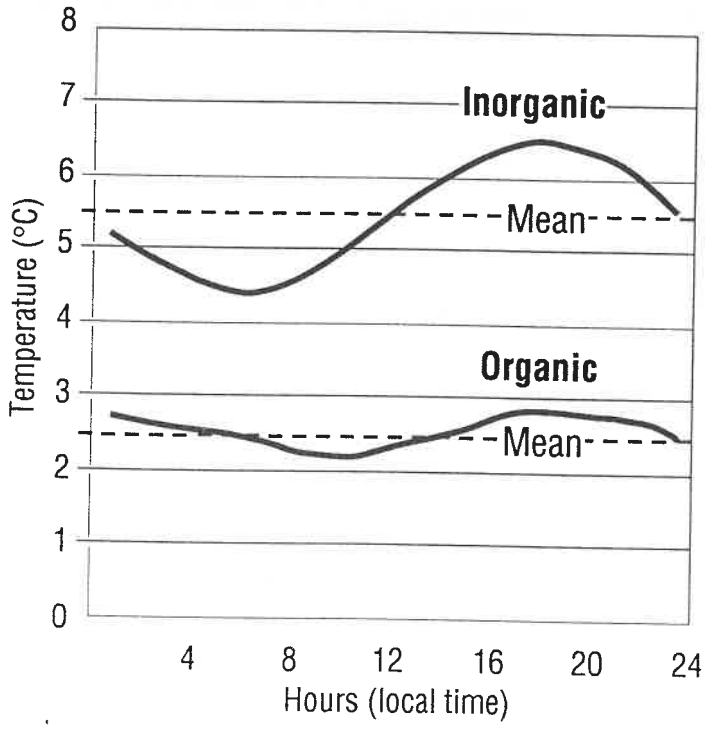


Figure 2.3. Mean diurnal 10-cm ground temperature regimes at adjacent organic and inorganic soils in the Mackenzie Delta (from Williams and Smith, 1989).

2.2. Biological characteristics

2.2.1. Vegetation history

The tundra as a biome is relatively young, having developed in the early Pleistocene (Bliss, 1981). However, the characteristic floras of tundra developed earlier, probably during Miocene-Pliocene, in the highlands of central Asia and in the Rocky Mointains of North America. In the same period (late Tertiary) present tundra areas were covered by various types of mixed and coniferous forests, now associated with the boreal forest region.

Circumpolar regions of tundra-steppe environments developed during the Pleistocene. Rapid expansion of these areas during this period may have been

associated with the success of such now extinct macrograzers as Mammoths (Bliss, 1981).

A relative mild climate with spruce forests north of their present limit occured in the last interglacial (Eem). During the last glaciation (Wisconsin) major parts of the northern Eurasian and American continents were covered by the Eurasian and Laurentide ice sheets respectively. There were, however, non-glaciated pockets extending far north as e.g. in parts of Beringia and northern Yukon Territory. In such non-glaciated areas the diversity of animal and plant species and the general development of soil and plant communities have been shown to be no more complex than comparable glaciated land masses nearby (Bliss, 1981). This has lent support for the theory that at present arctic ecosystems, including soils, are in equilibrium with the prevailing climate. Thus, they have been considered quite stable, although many areas have been deglaciated for only 3000 to 8000 years. A discussion of whether this is still the case, and what might threaten such stability, forms part of a later chapter.

2.2.2. Biogeographical subcategories

From a biological viewpoint the Arctic is often defined as the lands beyond the climatic limit of tree growth in upland habitats between river drainages. These areas have often, particularly in North America, been considered to consist of only two floristic units, tundra and polar desert. Furthermore, in the western hemisphere, the application of basic, diagnostic characteristics for the division of the Arctic into simple subdivisions based on the degree of closedness of the vegetation has been widespread (Aleksandrova, 1980). Traditionally, scientists of the former Soviet Union have to a larger extent than North Americans identified, a larger number of biogeographical subzones (Bliss and Matveyeva, 1992)

Andreev and Aleksandrova (1981), for example, identified 13 arctic vegetation types and five subzones within the tundra zone based on the species composition and characteristics related to life-form and migration history. In addition, they identified a number of longitudinal provinces across the Eurasian continent.

Part of the reason for the varying use of subzones by North American and Eurasian scientists is probably differences in the physical geography of the two continents. In Eurasia, most of the land north of the treeline is continental, with groups of islands in the Arctic Ocean. The climate shows a gradual northward shift over continuous land-masses, which provides the basis for major subzones or "belts" of vegetation. In contrast, the North American continent has a different geomorphology: no unbroken landmass extends to 78°N as in Eurasia and trans-longitudinal mountain ranges, sea-barriers and icecaps combined with their respective climatic influence cause a coarse-grained mosaic vegetation pattern rather than arrangements in belts or zones (Bliss and Matveyeva, 1992). Bliss and Matveyeva (1992) provide the most recent overview of tundra subcategorisations with areal estimates on a scale appropriate for this study. The following description is taken largely from their paper. For a detailed discussion of tundra subdivisions see Aleksandrova (1980).

Shrub tundras. In North America, shrub communities are dominated by *Betula nana* and various species of *Salix*. The ground cover includes *Carex* and *Eriophorum* spp, numerous dwarf shrubs, grasses, and forbs; and an abundance of lichens and mosses. The most common dwarf shrubs belong to the genus *Ledum*, *Vaccinium*, *Empetrum*, *Rubus*, *Arctostaphylos* and *Cassiope*. The most common mosses include species of *Hylocomium*, *Aulacomnium*, *Polytrichum* and, figuring prominently, *Sphagnum*. Important lichen families are *Cladina*, *Cetraria* and *Cladonia*. The tall shrub canopy is 40-60 cm high, and the heath shrubs and forbs are 10-20 cm in height; the cryptogams provide a more or less complete ground cover. This type of vegetation is widespread in Arctic Alaska and in many areas of Arctic Canada except for the central and eastern parts.

Shrub tundra in Eurasia extends on rolling uplands across much of Siberia. Again the taller shrubs are dominated by varying subspecies of *Betula nana* and species of *Salix*. The ground cover includes *Carex* spp. and heath (low-) shrubs, *Ledum*, *Vaccinium*, *Empetrum*, *Arctous*, *Dryas* and *Cassiope* spp. There is a continuous moss

cover comprising *Hylocomium*, *Tomenthypnum*, *Aulacomnium* and *Dicranum* spp. Lichens *Cladina*, *Cladonia* and *Cetraria* spp. occur within the mosses. The tallest shrub (up to 2 m), forming thickets are *Alnus fruticosa*, which are important in many places from Ural to Chukotka. Also, thickets of *Salix* spp., 1-2 m tall, occur mainly in drainages and along river banks (see section 2.1.2).

Tussock and sedge-dwarf shrub tundras. In the vast areas of tussock and sedge-dwarf shrub tundra, many of the shrubs mentioned above are still present, but they do not form a canopy and they occur almost exclusively in depressions, on raised polygons in mires or along river banks. The tussock and sedge-dwarf shrub tundra corresponds largely to what scientists of the former Soviet Union refer to as "typical tundra"(Chernov, 1985).

In the Siberian Arctic, the sedge-dwarf shrub tundra is the most widespread. Here the main species are mosses *Hylocomium*, *Tomenthypnum*, *Aulacomnium* spp. and species of lichens *Cladina* and *Cladonia*. Dwarf shrubs species of the genus *Vaccinium*, *Salix* and *Cassiope* are common as are *Carex*, *Ptilidium* and *Dryas* spp. in the ground layer. In North-western Siberia *Dryas octopetala* dominates many dry communities.

Tundras dominated by *Eriophorum vaginatum* tussocks with *Carex* spp., along with the common dwarf-shrub species and an abundance of mosses and lichens occupy large areas in western parts of North America particularly Alaska (Figure 2.4). Again dwarf shrubs of *Betula* and *Salix* are common along with numerous forbs, grasses, and an abundance of mosses, including species of *Sphagnum*, *Hylocomium*, *Dicranum*, *Aulacomnium* and *Tomenthypnum*. Common lichens include *Cetraria*, *Cladonia*, *Cladina* and *Thamnolia* spp. Tussock tundra is more limited in the eastern parts of Arctic America and West Siberia. However, large tracks of land dominated by *Eriophorum vaginatum*, along with the common heath shrubs and an abundance of lichens and mosses, are found in East Siberia and Chukotka.

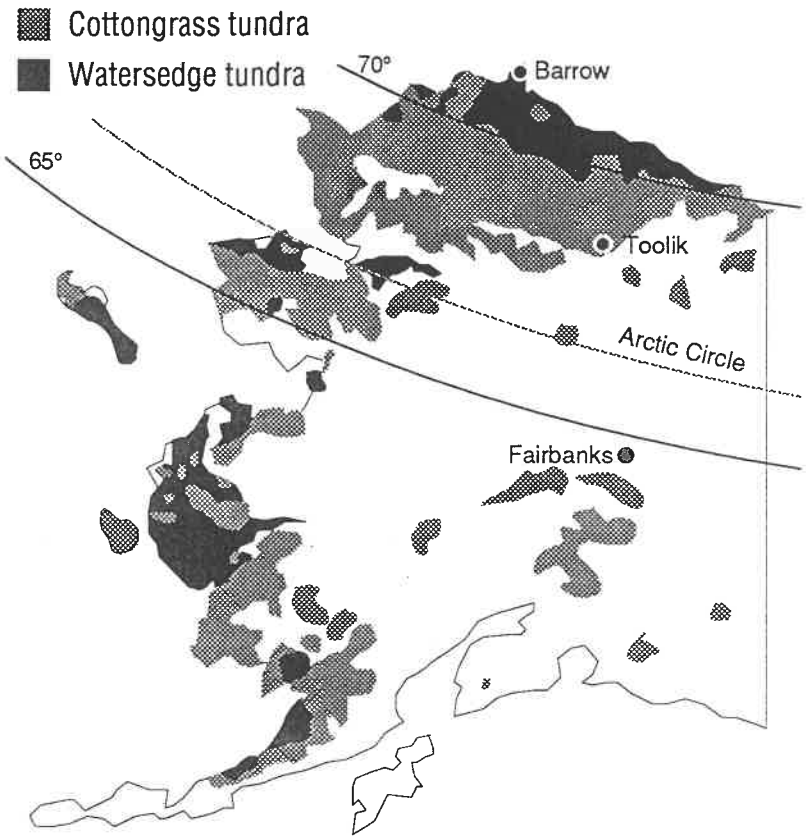


Figure 2.4. Distribution of tussock and wet sedge tundra in Alaska (modified from Oechel, 1989).

Wet tundras. Wetland plant communities in North America dominate on the Coastal Plain of Alaska and in the flat coastal areas in the Yukon. They extend on islands in the Mackenzie River Delta and eastward on the Tuktoyaktuk Peninsula. Peat in these communities reaches considerable depths typically, 1-5 m. Only a shallow (20-40 cm) active layer develops in these cold, wet, soils (see below).

The dominant sedges are in particular species of *Carex* and *Eriophorum.* Grasses include *Arctagrostis*, *Dupontia* and *Arctophila* spp.. Mosses are abundant including species of *Aulacomnium*, *Calliergon*, *Ditrichum*, *Drepanocladus*, *Hylocomium*, *Meesia*, *Tomenthypnum*, and *Sphagnum*. Various dwarf-shrubs are common along the rims of the mires and on raised hillocks. Sedge-dominated mires occur in lowlands across the Canadian Shield but in limited extent.

Wet tundras form a prominent part of the Siberian tundra. They are especially well developed in the central part of the Yamal Peninsula and in the lowlands of the Yana, Indigirka, and Kolyma river basins. Various types of mires have been identified in the Russian Arctic but all are dominated by species of *Carex*, *Eriophorum*, *Caltha* and *Comarum* as well as moss species *Drepanocladus*, *Meesia*, *Calliergon*, *Polytrichum*, *Sphagnum* and *Cinclidium*. Dense thickets of *Betula nana* are typical for large peat hillocks up to 30 m in diameter. *Dryas*, *Vaccinium*, *Salix* and *Betula* spp. typically grow on the rims of polygonal mires.

Mires similar to the above but without the same abundance of shrub species, *Betula*, *Salix* and *Vaccinium* are also common in high arctic communities particularly in the Northwest Territories of Canada.

Polar semideserts. The tundras described so far, except for mires, are exclusive to what Bliss and Matveyeva (1992) identify as the Low Arctic. The polar semideserts are in the High Arctic which is exemplified by a number of structural and floristic changes in vegetation. These changes includes a shift from the predominance of low-shrub (*Betula*, *Salix*), dwarf-shrub (heath species), and cottongrass-tussock-dwarf shrub tundras to an

open vegetation dominated by cushion plants (*Dryas* and *Saxifraga* spp.), prostrate shrubs of *Salix arctica*, and rosette species of *Saxifraga*, *Draba* and *Minuartia*.

Along a south-north transect in the continental Siberian true tundra, the boreal elements of the low shrub species are the first to disappear. Subsequently other dwarf shrub species decrease significantly in the transition to high arctic tundra. Arriving on the "true tundra", as named by Russian scientists, the dominating species are *Salix polaris*, *S. arctica*, the graminoids *Alopecurus alpinus*, *Deschampsia borealis* and *Luzula confusa*. Comparable tundras in North America have been called polar semideserts.

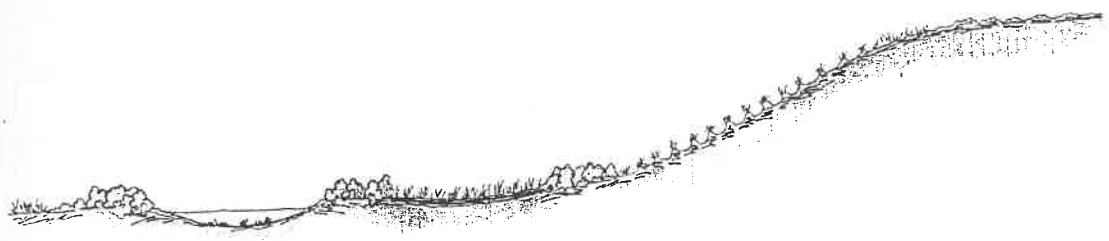


Figure 2.5. Toposequence of typical rolling tundra landscape in Alaska (see text). Drawing by Ann-Britt Broström.

As mentioned earlier, the above major vegetation types tend to appear in zones in the Eurasian Arctic and in a mosaic in North America. However, all over the arctic tundra on a local scale these vegetation types are mixed, depending on local climatic, hydrologic and topographic features. For example, a typical local toposequence from a tussock tundra "typical" environment on the North Slope of Alaska often looks like that

illustrated in Figure 2.5. Along the river, thickets of shrubs are formed due to the absent permafrost. In the low-lying area at the valley bottom adjacent to the river an area of wet sedge tundra has developed, surrounded by dwarf shrubs. Upslope from this is the tussock tundra environment, covering most of the hillside but degrading into polar semidesert vegetation on the wind-exposed hilltop. This transition, occuring within a few hundred metres or less, is a common feature of tundra environments in landscapes with rolling hills.

Table 2.4 shows estimated areal extents of the described vegetation types.

Table 2.4. Areal extent (x 10^{12} m^2) of various tundra types in different regions of the Arctic (from Bliss and Matveyeva, 1992).

Vegetation type	Alaska	Canada	Greenland, Iceland	Eurasia	Total area
Low Arctic					
tall shrub	0.018	0.026	0.018	0.112	0.174
low shub	0.090	0.264	0.032	0.896	1.282
tussock, sedge-shrub	0.126	0.088	0.036	0.672	0.922
wet sedge	0.104	0.176	0.040	0.560	0.880
semidesert	0.018	0.326	0.014	-	0.358
ice caps	-	-	0.776	-	0.776
High Arctic					
wet sedge	0.004	0.096	-	0.032	0.132
semidesert	-	0.720	0.093	0.192	1.005
polar desert	-	0.640	0.127	0.080	0.847
ice caps	-	0.144	1.031	0.016	1.191
Total land	0.360	2.336	0.368	2.544	5.600
Total land plus ice caps	0.360	2.480	2.167	2.560	7.567

2.3. Tundra soil formation and characteristics

The tundra soils are young, typically dating back less than 12,000 years (Everett *et al.*, 1981) except in some areas that were not ice-covered during the latest glaciation. In some local sectors the landscape has been ice-free only during relatively recent time. In addition to disappearance of the ice cover itself, major areas of the Arctic have undergone isostatic adjustment resulting in emerging landforms (Linell and Tedrow, 1981). A large area surrounding Hudson Bay and extending northward to the arctic islands was depressed well below sea level during Pleistocene, as was northern Scandinavia and the lower courses of major rivers of Siberia. Some emerged landforms have accumulated organic-rich sediments, which, coupled with flat terrain and poorly developed drainage patterns, have resulted in sluggish surface drainage and formation of extensive wetland conditions. The presence of permafrost further restricts soil and plant development. Consequently, decomposition, release of nutrients, and synthesis of secondary minerals from weathering of clay all progress very slowly.

Classification of arctic soils are still quite controversial with different systems being used in the U.S., Canada, and Russia respectively. A discussion of the different approaches lies beyond the scope of this thesis. For a detailed discussion of arctic soil classification see Tedrow, 1977; Linell and Tedrow, 1981. This section will deal more with the physical characteristics.

Figure 2.6 shows an idealised genetic sequence of major soil groupings along a moisture gradient. The process of podzolization is limited to well-drained soils with a deep active layer. Where dwarf shrub species predominate, weakly developed podzols (Spodosols) are found. Less well developed soils of uplands and dry ridges are the arctic brown soils (Inceptisols). The most common group of soils in the Low Arctic region includes the tundra soils (Inceptisols) underlying cottongrass-dwarf shrub and some sedge communities of imperfectly drained habitats. These soils form under the process of gleization. Poorly drained lowlands where soils remain saturated all summer

accumulate peat. These wet and highly organic tundra soils belong to the Histosol group.

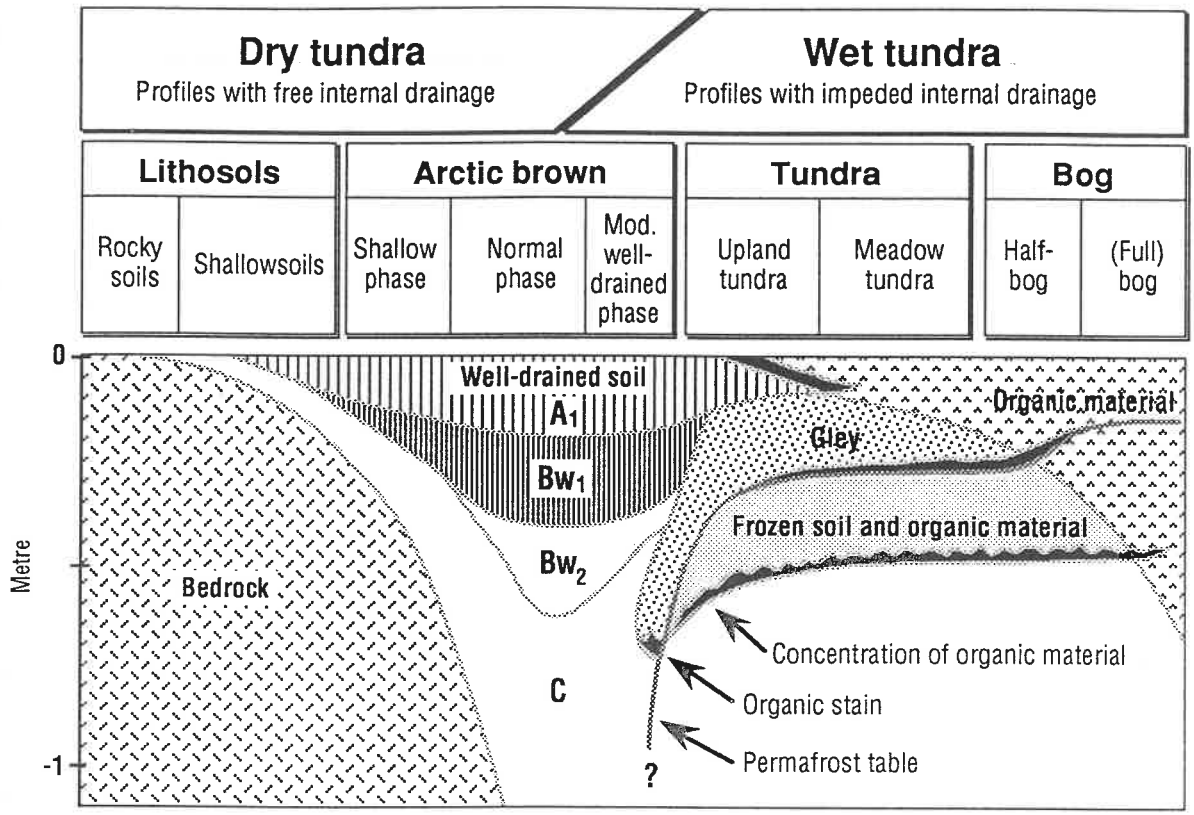


Figure 2.6. Generalised diagram of major soils in the Arctic (modified from Tedrow, 1977).

The drier arctic soils (left part of Figure 2.6), podzols and arctic brown soils show some translocation of humus and iron, with iron-enriched B_2 horizons and weakly eluviated A_2 horizons in the podsols. Surface layers tend to be acidic (pH 6-4) and low in available nutrients, but quite well drained above the permafrost. Inceptisols (arctic tundra soils) are less well drained but show generally similar acid pH characteristics and low nutrient availability. They contain B horizons that have subangular to angular structures, are grayish in color, and include iron oxide mottles. Histosols of poorly drained lands are acidic (pH 6.5-5.0) and are similar to arctic tundra soils in having limited translocation of minerals into the B horizon.

2.4. Global geographical databases

The areal estimates in Table 2.4 are derived using ordinary maps. Digital mapping has an increasing importance for global change studies. The following will form a short discussion of tundra in relation to global digital databases.

All studies of trace gas flux using areal data for the global coverage of major biomes are quite recent. Early studies of global tundra methane flux (Ehhalt, 1974; Svensson, 1976) used Whittaker's (1971) figure for global tundra, including alpine tundra, developed for estimation of global NPP and plant biomass per ecosystem type. This areal figure is produced on the basis of conventional map sources and quotes a global tundra area of 8×10^{12} m^2.

Digital global vegetation and land use databases are important for studies relating to climatic change in that organization, modification and incorporation of geographical data for different purposes are made possible. Mathews (1983) produced the first such data base, using the UNESCO clasification system (UNESCO, 1973) for digitizing present global vegetation coverage at a 1° latitude by 1° longitude resolution. The UNESCO system classifies vegetation on the basis of lifeform, density, and seasonality, with supplementary terms on altitude, climate and vegetation architecture. Vegetation types are designated by a series of numbers and letters indicating, in order of increasing detail; formation class, formation subclass, formation group, formation, and subformation. Technically, at least 225 vegetation types can be designated with this system. However, Mathews used only 148 for her study and gave no quantitative information on subcategories of tundra (Mathews, 1983). Total tundra was estimated to cover 7.36×10^{12} m^2, much in agreement with the estimate based on conventional sources mentioned above, but less so with the estimate discussed in section 2.2.2.

Figure 2.7 illustrates how according to the database the tree line, the forest-tundra boundary, shows different geographical and vegetational characteristics in the eastern and western hemispheres. In Canada the boundary (~50°N) penetrates about 2000 km south of the position of the Eurasian boundary (~70°N) and gradational woodlands between the forest and tundra in Canada are a much less significant feature than those

in Eurasia (Mathews, 1983). The latter might partly be an artifact arising from different classification systems in the original conventional data sources used by Mathews. However, since the main limitation on tree growth is the summer temperature (rather than winter or annual mean temperature, see above) the stronger continentality of the Eurasian continent with warmer summer temperatures probably accounts for a real difference reflected in Mathews study (Figure 2.7).

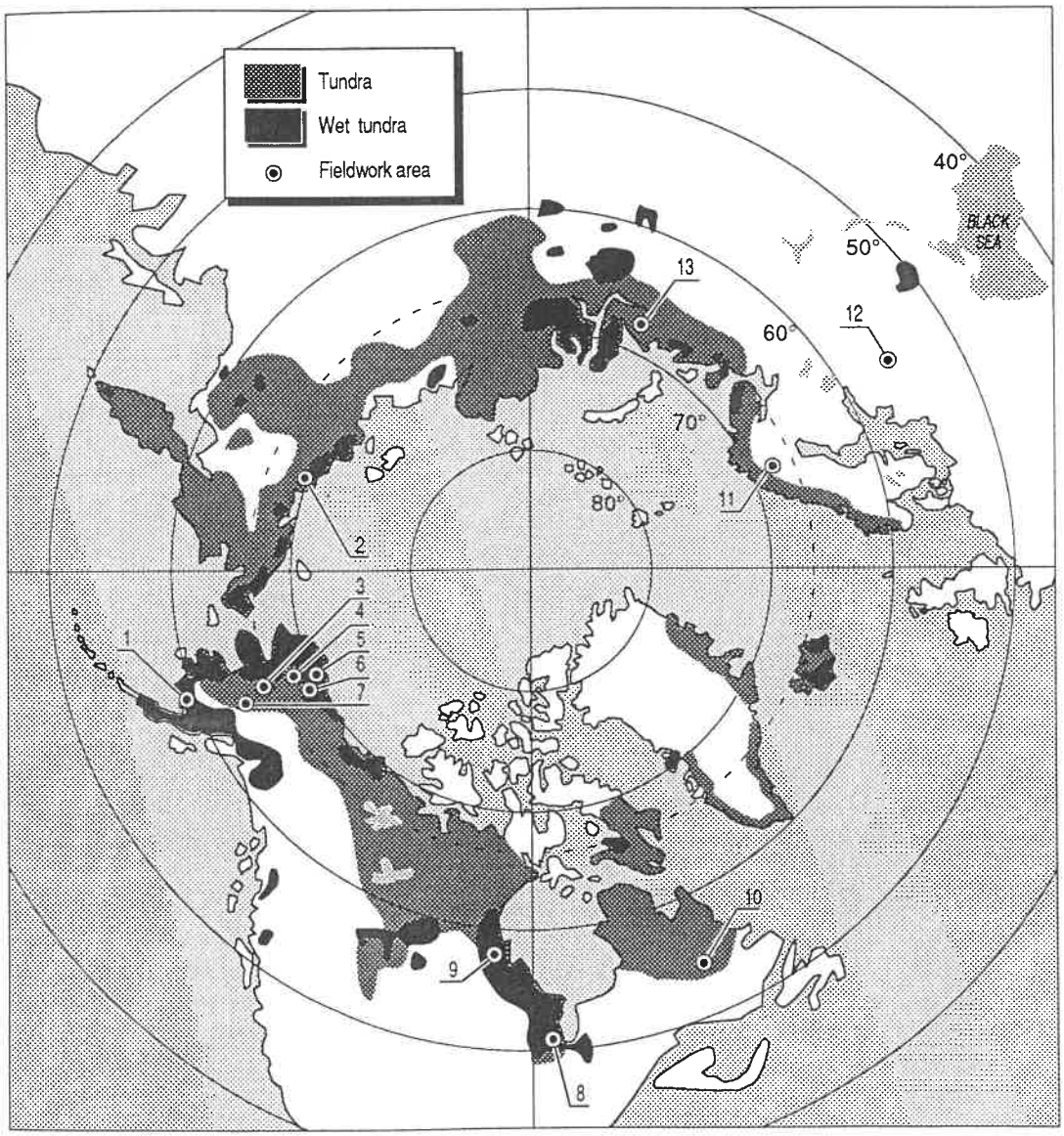


Figure 2.7. Global tundra distribution as derived from manual interpretation of digital maps produced on the basis of global databases. Numbers refer to field surveys of methane flux discussed later (see section 4.4.3).

As will be apparent from later chapters, soil wetness is of paramount importance when studying soil trace gas flux. Geographical information on the areal coverage of inundated versus varying stages of drier areas is needed when estimating global trace gas fluxes. As mentioned above, Mathews (1983) did not subcategorise tundra but in a later study of global methane emissions from natural wetlands, partly relying on the described data base, Mathews and Fung (1987) developed a further data base of global wetlands also at 1° resolution. It integrates three independent data sources to arrive at a global estimate of the area, location, and environmental characteristics of wetlands. The first is the above mentioned UNESCO based vegetation data base. The second is a global 1° resolution data base of soil properties digitized from FAO soil maps, from which a distribution of ponded soils were derived. The third is a global 1° resolution inundation data base compiled from a series of Operation Navigation Charts for pilots (ONC maps). The three data sets were combined for purposes of completeness and corroboration (Mathews and Fung, 1987).

Total wetlands north of 50°N were estimated at 2.7 x 10^{12} m^2 comprising of forested bogs, 1.723 x 10^{12} m^2; nonforested bogs, 0.884 x 10^{12} m^2; forested swamps, 0.03 x 10^{12} m^2 and nonforested swamps, 0.066 x 10^{12} m^2. Nonforested bogs represent the wet areas of the tundra environment: Subtracting these areas from the total tundra estimate of 7.36 x 10^{12} m^2 provides a rough subcategorisation of tundra based on degree of wetness (Figure 2.7). Thus, 6.476 x 10^{12} m^2 remain as covered by dry/moist tundra. This is obviously a very crude estimate and there are major differences between this and areal estimates developed for specific tundra studies. In these studies total tundra only amounts to 5.7 x 10^{12} m^2 compared to 7.36 x 10^{12} m^2 derived above from Mathews database (Mathews, 1983). The wet tundra corresponds well with the wetlands database estimating 1 x 10^{12} m^2 compared to 0.88 x 10^{12} m^2. However, this leaves an even larger difference in the figures for dry/moist tundra (shrub, tussock, sedge, dwarf-shrub tundras and polar semidesert), of 4.7 x 10^{12} m^2 (including polar desert) in Table 2.4 against the database derived 6.48 x 10^{12} m^2. This difference serves

to underline a general uncertainty factor that still exists in all global extrapolations of tundra ecosystem features and functions.

The crude subdivisions between wet and dry/moist tundra are the most frequently used scheme for extrapolating tundra methane fluxes (Whalen and Reeburgh, 1988; Bartlett *et al.*, 1992; Whalen and Reeburgh, 1992; Christensen, 1993) and for comparative reasons is used also later in this thesis.

In a digitised wetland data base based on regional map and litterature information, Aselman and Crutzen (1989) estimated a total wetland area north of 50°N (3.056 x 10^{12} m^2) larger than that shown by Mathews and Fung (2.712 x 10^{12} m^2). However, in their classification they did not distinguish between forested and nonforested wetlands. It is therefore not possible to use their data for direct comparison of the wet tundra estimate derived from Mathews and Fung. The differences in areal estimates does, however, indicate that there are still major uncertainties in the areal coverage estimates of explicitly wet tundra, and indeed the gradient of wetness within the "rest" category.

A higher wetness class resolution within the tundra is needed, as well as information on seasonal cycles in degree of wetness of tundra soils on a circumpolar scale. Recent progress in detecting spatial and temporal distribution of freeze-thaw events in tundra soils by remote sensing provides hope for such information to be available in digital form within the forseeable future (Villasenor *et al.*, 1993).

2.4 Summary

This chapter provides an overview of physical and biological tundra characteristics; the main issues being the interaction between climate and the soil environment, in combination with associated biological processes.

Net radiation in combination with albedo effect have been identified as the main controlling factors for plant life in the Arctic. The main limitation on plant growth is the length of a summer growing season rather than annual mean temperature or low temperatures in winter. Fairly good approximations of the distribution of major

vegetational groups (e.g. the tree-line) can be made from "isolines" of a certain number of degree days above 0°C or the mean July temperature isotherm.

The tundra soil environment is generally wet which is due to low rates of evaporation and poor drainage rather than high precipitation. Poor drainage is consequential on the presence of permafrost, one of the features that makes the tundra different from other seemingly similar environments. The stability of permafrost is dependent upon a mean annual surface temperature below 0°C; the depth of the seasonally thawed layer (the active layer), where most of the processes of relevance to this thesis take place, is determined by the interaction of climate, soil heat capacity and insulating effects of, for example, vegetation and peat. In short, the unique tundra soil environment is primarily controlled by climate, although the relationship is complex involving many interdependent processes.

The tundra biome is young, having developed since the last glaciation. So too are the soils, but they have nevertheless accumulated considerable amounts of carbon (see Chapter 3) and large parts of the tundra consist of wet organic Histosols.

Vegetation subcategories have been used extensively by Russian scientists while the North Americans traditionally have had more crude characterisations. The most detailed information on the areal extent of tundra operates with six subcategories: polar desert (0.8×10^{12} m^2), semi-desert (1.5×10^{12} m^2), wet sedge (1.0×10^{12} m^2), tussock (0.9×10^{12} m^2), low shrub (1.28×10^{12} m^2) and tall shrub (0.23×10^{12} m^2) yielding a total of 5.71×10^{12} m^2. This estimate, developed for the specific purpose of studying tundra, shows some disagreement with a global digital vegetation database derived figure for global tundra of 7.36×10^{12} m^2.

Chapter 3

Carbon cycling in arctic tundra

3.1. Net primary production and total decomposition

A widely cited figure for mean tundra net primary productivity (NPP) is 65 g C/m^2/yr (Whittaker, 1975; Schlesinger, 1991). This figure is lower than for most other global biomes due to the low annual net radiation in arctic areas (the estimate for tropical rain forest is 900 g C/m^2/yr). However, it is a mean of annual totals, and covers relative high rates of NPP found on the tundra during the short growing season. Large differences between different tundra plant communities are also prominent depending on light, temperature, nutrient and moisture factors controlling production. NPP ranges from 14-30 g C/m^2/yr for semidesert/heath tundra, over 90-150 g C/m^2/yr for tussock tundra, to >300 g C/m^2/yr in shrub tundra (Shaver and Chapin, 1991; Oechel and Billings, 1992) (Figure 3.1). Rates as high as 1000 g C/m^2/yr have been estimated at shrub tundra sites (Heal *et al.*, 1981).

Globally about 100 x 10^{15} g C enters terrestrial vegetation every year as gross primary production. About half of this is lost through plant respiration yielding a global terrestrial NPP of approximately 50 x 10^{15} g C/yr. The tundra biome contributes only about 1% or 0.5 x 10^{15} g C to this figure (Whittaker, 1975; Schlesinger, 1991). Tundra and boreal wetland ecosystems have, however, provided a significant sink for atmospheric carbon through the Holocene due to low decomposition relative to plant production (Adams *et al.*, 1990). There is, though, controversy as to the size of the present tundra carbon pool and also the present carbon balance of tundra.

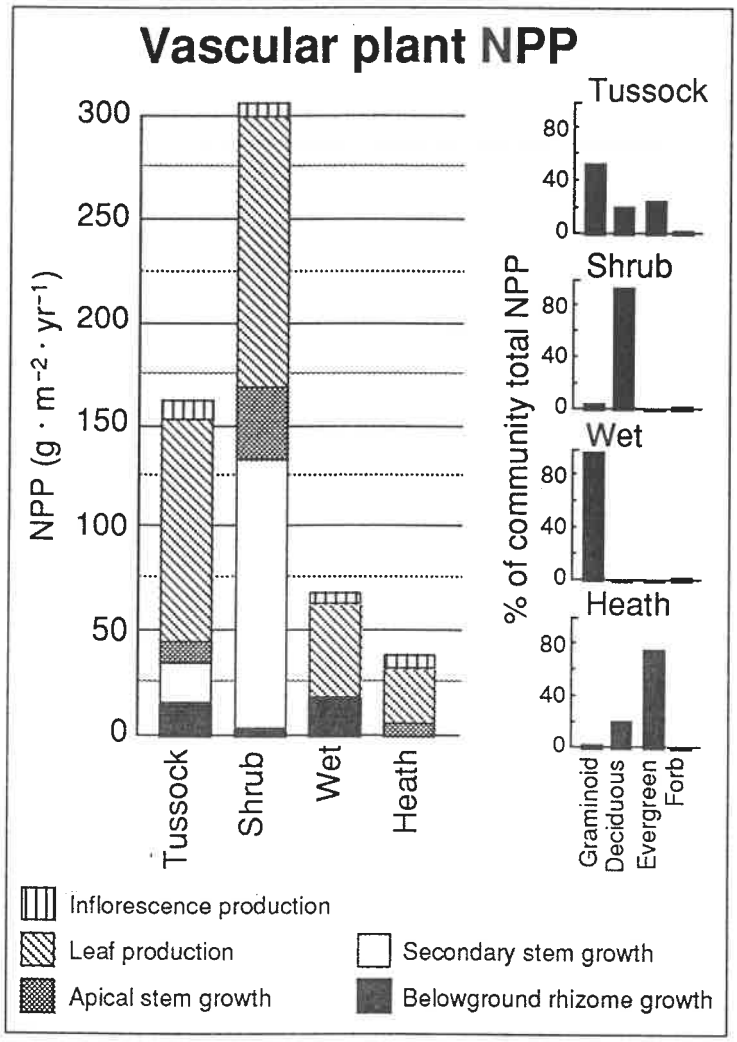


Figure 3.1. Net primary productivity (NPP) by vascular plants for each of four tundra types in Northern Alaska (modified from Shaver and Chapin, 1991).

Table 3.1. Areal distribution and carbon accumulation of tundra ecosystems (modified after Oechel and Billings, 1992).

	Polar desert	Semi desert	Wet sedge	Tussock	Low shrub	Tall shrub	Total
Area (x $10^{12}/m^2$)	0.8	1.5	1.00	0.90	1.28	0.23	5.71
Standing crop (kg C/m^2)							
Biomass (above + belowground)	0.002	0.29	0.95	3.33	0.77	2.61	
Dead organic matter[b]	0.091	7.2	13.4	29.0	3.8	0.4	
Total	0.093	7.49	14.35	32.33	4.57	3.01	10.7[a]
World total (x 10^{15} g)							
Biomass	0.0	0.43	0.95	3.00	0.98	0.60	5.84
Dead organic matter[b]	0.07	10.8	13.4	26.1	4.86	0.09	55.13
Total	0.07	11.23	14.35	29.10	5.85	0.69	61.06

[a] Areally weighted mean

b These values include only A and organic horizons. Enmixed organic material is probably significant in wet sedge and tussock tundra to about 20% of the stated values, bringing the world tundra total to 68.96 x 10^{15} g. Organic carbon permanently frozen in permafrost is not included.

Carbon storage. Older literature values for mean carbon stocks in tundra are usually around 20 kg C/m^2 (Schlesinger, 1977; Post *et al.*, 1982). Post *et al.* (1982) estimated global carbon pools based on soil samples from varies lifezones. The tundra samples were from wet and moist tundra which may have given a bias towards higher carbon content. More recent estimates using an areal weighted mean of tundra carbon storage based on data for six tundra vegetation types, result in smaller figures of around 11 kg C/m^2 (Table 3.1) (Miller *et al.*, 1983; Bliss and Matveyeva, 1992; Oechel and Billings, 1992). Although these studies agree on the overall mean, Bliss and Matveyeva show surprisingly large disagreement with the others on where the bulk of this carbon is located. According to Miller *et al.* (1983) and Oechel and Billings (1992) the largest amount of carbon is stored in tussock tundra environments followed by wet tundra and semidesert. Of total tundra carbon these studies estimate that almost 50% is stored in tussock tundra and less than 25% in wet tundra. This is in sharp contrast to Bliss and Matveyeva (1992) who estimate by far the largest proportion of total tundra carbon in the wet tundra (70%) and only 13% in tussock environments. The estimates for wet tundra carbon in all studies are conservative since only the top soil layer (the A horizon) is included in the source material (Chapin *et al.*, 1980; Bliss and Richards, 1982; Miller *et al.*, 1983). The difference between the two estimates might therefore not be significant in terms of wet tundra. However, in terms of tussock tundra Bliss and Matveyeva show an unexplained difference to the others, particularly between the estimates for soil organic carbon. Bliss and Matveyeva estimated 1 kg C/m^2 while Miller *et al.* (1983) and Oechel and Billings estimated 29 kg C/m^2. Most data in the literature seem to support the latter, higher, estimate. Table 3.1 shows estimates for carbon stocks thought most likely to be correct.

So there is controversy about present carbon stocks. However, large stocks of organic material have undoubtedly accumulated in tundra environments and production must therefore have exceeded decomposition at some time in the past. Recent estimates (Gorham, 1991; Marion and Oechel, 1993) indicate that northern ecosystems as a whole

still constitute a small net sink for atmospheric carbon (between 0.03 and 0.3 Pg/yr). Current accumulation rates are difficult to assess. Because the rates vary with conditions and ecosystem type, soil carbon accumulation is positive in some areas and negative in others. The overall balance is still uncertain (Oechel and Billings, 1992).

Wet tundra environments are probably still accumulating carbon but the estimated rates varies substantially. Chapin *et al.* (1980) found relatively high carbon accumulation rates of 109 g C/m^2/yr at Point Barrow. Clymo (1984) notes how simple accumulation rates calculated on the basis of two ^{14}C measurements divided by the depth between them might be seriously misleading. The validity of such data depends on whether the profile of bulk density is integrated in the calculation (Clymo, 1984). Incorporating such data Marion and Oechel (1993) quotes figures for carbon accumulation in organic soil in the Sub- and Low Arctic of 9-11 g C m^{-2} yr^{-1} increasing to 23-35 g C m^{-2} yr^{-1} in boreal peatlands. Disregarding the uncertainties about present wet tundra carbon accumulation estimates, this ecosystem has shown potential for long-term sequestration of carbon, and is probably still accumulating.

Carbon accumulation in tussock tundra is a more complex question. Miller *et al.* (1984) estimated carbon accumulation rates in Alaskan tussock tundra between 50 and 70 g C/m^2/yr. Recent measurements from the same region indicate, however, that tussock tundra is actually losing carbon at rates of 53 to 286 g C/m^2/yr (Grulke *et al.*, 1990; Oechel *et al.*, 1993). An additional approximately 20% loss could be associated with transport of carbon to lakes and streams and subsequent release to the atmosphere (Kling *et al.*, 1991). These figures may look very dramatic especially considering, as shown above, total carbon content in tussock tundra has been estimated as low as 1 kg/m^2. Marion and Oechel (1993) acknowledge that these large negative carbon accumulation rates probably are short term phenomena and that tundra ecosystems in general over a longer time-scale (centuries to millenia) will continue acting as atmospheric carbon sinks at rates estimated at 0.03-0.07 Gt/yr.

Not all the stored carbon might has the same potential for decay. Hogg *et al.* (1992) and Hogg (1993) have shown considerable differences in decay potential of peat

from different depths in Canadian and Swedish mires respectively. In a Canadian boreal mire they found 4-9 times higher potential for CO_2 release under similar aerobic conditions from the 0-10 cm layer than from the 30-40 cm layer (Hogg *et al.*, 1992). Similar studies are needed in tundra environments, to assess possible total carbon loss from these environments following global warming.

All carbon mineralisation leads to production of CO_2 and/or CH_4 and the present and potential emission of these gases, in particular CH_4, are central questions in this thesis.

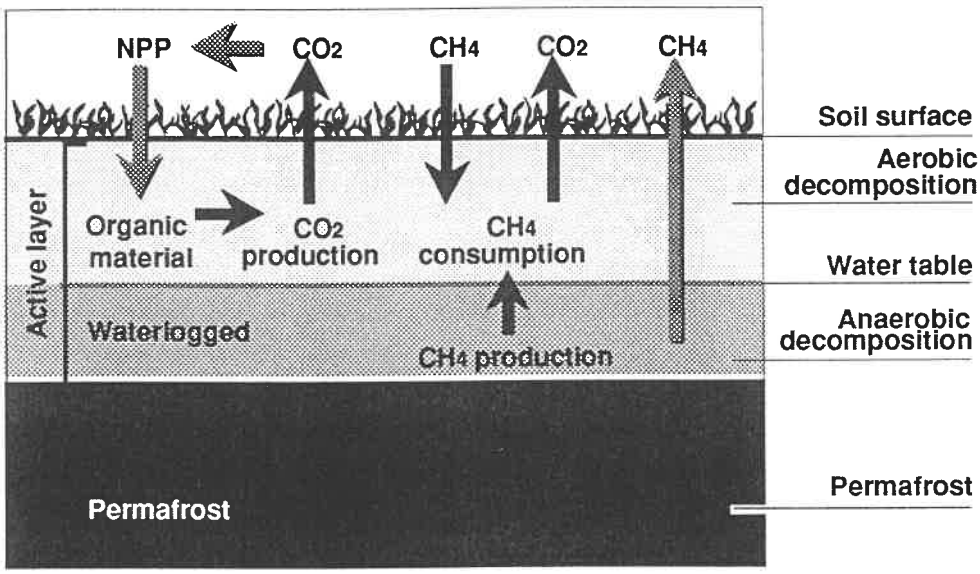


Figure 3.2. Simplified diagram showing decomposition in a wet tundra environment.

3.2. Soil decomposition

A generalised picture of decomposition in a wet tundra environment is shown in Figure 3.2. The grey arrows surrounding NPP reflects the surplus over total decomposition which causes build-up of organic material in the ground. The water table forms a general boundary between aerobic and anaerobic soil environments allthough anaerobic microenvironments are known to occur in smaller pores above the water table. In mire ecology the aerobic zone is normally referred to as the "acrotelm", characterised by high hydraulic conductivity and relative fast decay rates. The anaerobic zone is named

the "catotelm" characterised by low hydraulic conductivity, reducing conditions, and low rates of decay (Clymo, 1984). The widespread anaerobic environment causes a surplus of methane production over consumption which, in Figure 3.2, also is illustrated by a grey arrow.

The plant composition of tundra subcategories gives rise to differences in chemical composition of the litter. Wet tundra dominated by monocotyledons produces remains that are low in lignin (< 25%) and have moderate concentrations of soluble constituents (< 25%) and mineral nutrients (C:N 30-50). Shrub tundra environments have a much higher fraction of woody tissue, with lignin concentrations of 25-40%, a low soluble fraction (< 10%), and low nutrient concentration (C:N 60-120) (Heal *et al.*, 1981). Mosses, a major component of tundra vegetation (see section 2.2.2), also have low nutrient concentrations. In drier areas lichens provide substantial amounts of material for decomposition which are low in lignin and cellulose, but high in hemicellulose, and thus may require different decomposer organisms (Heal *et al.*, 1981).

Table 3.2. Numbers of heterotrophic bacteria at Moor House IBP tundra biome site producing colonies on plates incubated aerobically and anaerobically (Holding, 1981).

	Bacterial count		
Horizon	Aerobic plates	Anaerobic plates	Ratio aerobic/anaer.
Litter	260±66	5.9	44
Black-brown	110±15	9.3	12
Green-brown	76±19	6.3	12
Red-brown	15±4	0.5	30

Taxonomically, filamentous fungi, yeasts and bacteria of tundra soils show small differences from populations found in other regions, and no types unique to tundra have been recognised (Holding, 1981). Microbial populations taking part in decomposition of soil carbon are largest in the aerobic upper soil layers (Table 3.2). In general few

percent of the carbon released from tundra soils is the result of anaerobic processes leading to release of CH_4 (see section 7.3). This proportion varies a lot though, depending on soil wetness. About 50% of the CO_2 produced from aerobic microbial decomposition is a result of fungal respiration with the remainder primarily being bacterial (Holding, 1981). The total output of CO_2 is a result not only of microbial decomposition but also of root respiration and soil fauna activity. Few data on the relative proportion of these processes exists. However, microbial decomposition and root respiration are probably the most significant processes, contributing about equally to soil respiration (Heal *et al.*, 1981).

Total soil decomposition in tundra environments is low compared with other biomes, reflecting the cold, wet and general unfavourable conditions which promotes the build-up of organic material in tundra soils discussed above.

3.3. Methane biogeochemistry

In anaerobic environments the most significant process leading to mineralisation of organic material is carried out by methanogenic - methane producing - bacteria. This is a group of microorganisms which, along with the extremely halophilic and the thermoacidophilic bacteria, forms a distinct biological kingdom known as Archaebacteria (Oremland, 1988). Halophilic bacteria inhabit brines (water saturated with salt) while thermoacidophilic live in hot springs, highly acidic soils, and even near volcanic areas at the ocean floor. Methanogens are strict anaerobes and require highly reducing conditions (Eh $\leq$ -300 mV) for growth. Consequently, they proliferate in aquatic sediments, inundated soils, animal gastrointestinal tracts and sewage. Like the former mentioned groups they occur also in high temperature, hypersaline, and extreme pH environments, and they are well adapted to cold environments like tundra soils (Svensson, 1984). Viable methanogens have even been reported in deep permafrost at year-round temperatures of approximately -10 °C (D. Gilichinsky, pers. comm., 1992).

Methanogenic bacteria can metabolize only a restricted number of compounds which provide energy for their growth. Recognized substrates include hydrogen

reduction of carbon dioxide, acetate, formate, methanol, methylated amines, carbon monoxide, and dimethyl sulfide (Oremland, 1988). In cold, relatively acid, soils (like most tundra soils) reduction of carbon dioxide and acetate fermentation are main pathways of methane production; Svensson (1984) found the latter to be the dominant process in an acid sub-arctic mire.

Methanogens rely on other microorganisms to provide them with substrates. The breakdown of organic matter in anoxic ecosystems is a very complex process generally referred to as an "anaerobic food web" rather than a simpler food chain (Cicerone and Oremland, 1988). A variety of non-methanogenic anaerobic microbes attack complex organic compounds, ultimately producing the methanogenic substrates (Figure 3.3). Studies of the response of methane production to changing environmental conditions should therefore, from a microbiological viewpoint not only consider single factors, as for example temperature, since the substrate for methanogens might be limited by the activity of other microbes that could have a different response to environmental variables.

Activity of methanogenic bacteria is influenced also by competition for substrate, particularly that imposed by sulphate-reducing bacteria. If present, sulphate-reducing bacteria will normally outcompete methanogenic bacteria by having higher affinity for the two most important methane precursors, hydrogen and acetate. Thus in environments which contain abundant sulfate, such as marine and coastal sediments, methanogenesis usually takes place beneath the zone of sulfate depletion (Oremland, 1988). Methane emissions from coastal wetlands are therefore generally lower than from comparable environments further inland (Bartlett *et al.*, 1985).

Microbial methane consumption is a very important control on net methane emission from soils. In view of the widespread occurence of reducing conditions, early studies of tundra soil organic matter decomposition found it surprising that more substantial methane emission had not been reported, and it was suggested that methane-oxidizing bacteria could provide a significant role in limiting flux to the atmosphere (Heal *et al.*, 1981). It has now been shown that methane oxidisers in soils comparable to

northern wetlands and tundra are capable of consuming more than 90% of the methane produced at depth (King, 1990; Oremland and Culbertson, 1992) and it has been estimated that about 55% of the CH_4 produced in global tundra soils is oxidised before entering the atmosphere (Reeburgh *et al.*, 1993). Methane-oxidizers in sub-arctic tundra soils have also been shown to consume methane at sub-ambient levels indicating the potential for tundra soils to act as atmospheric methane sinks as well as sources (Whalen and Reeburgh, 1990a). Direct uptake and consumption of atmospheric methane on relatively dry soils have been observed by various authors (see review by e.g. Wahlen, 1993). It has been estimated that this sink globally amounts to about 1-15% of the chemical sink in the atmosphere (see section 3.5).

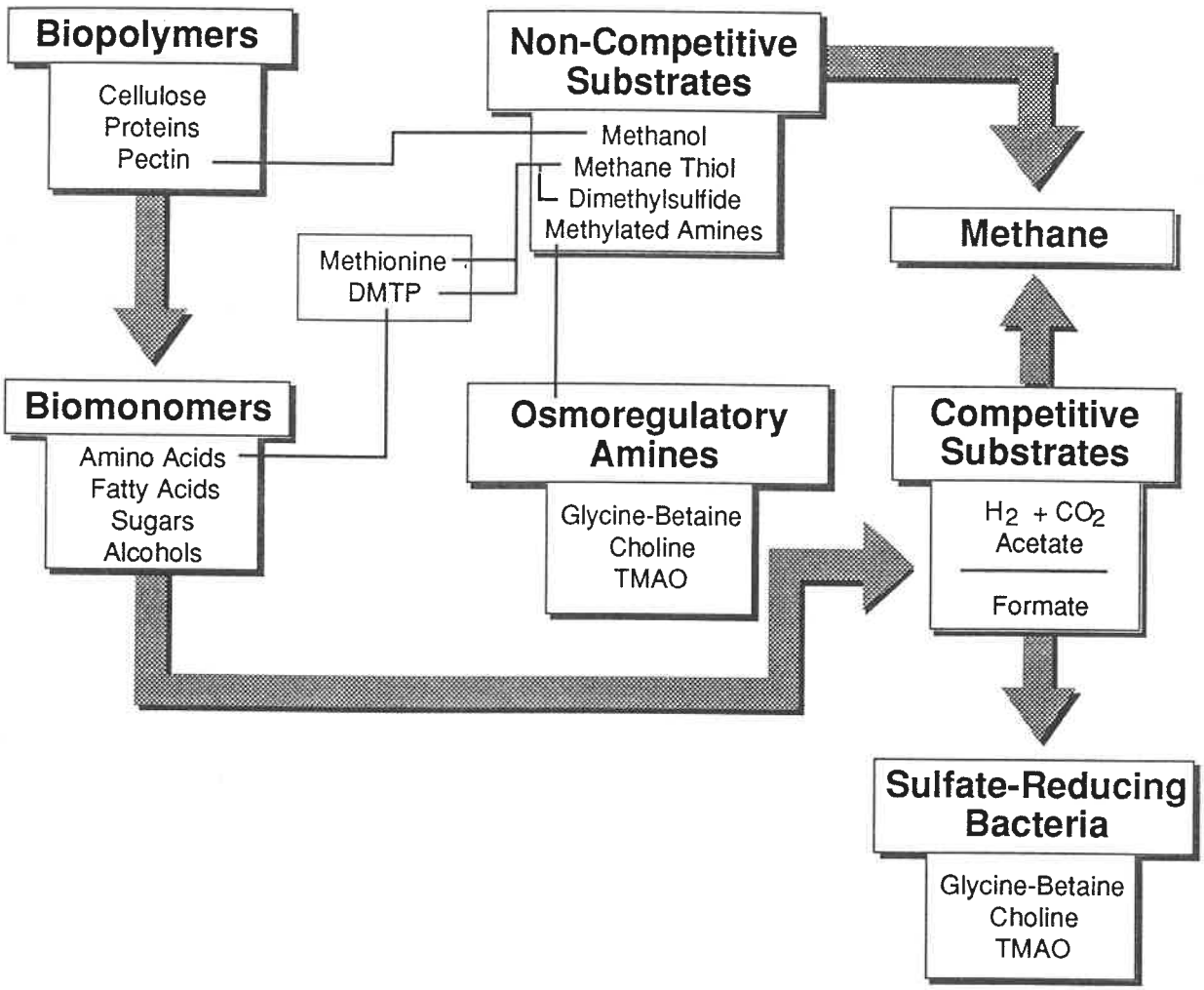


Figure 3.3. Anaerobic foodweb for microbial ecosystems (modified from Oremland, 1988).

Methane can be oxidised by both aerobic and anaerobic bacteria. Anaerobic methane oxidation is as yet a poorly understood process that seems quantitatively to be of most significance in marine environments (Alperin and Reeburgh, 1985). It will therefore not be further considered here.

Aerobic methane-oxidizing (methanotrophic) bacteria are eubacteria which grow only on methane and/or other one carbon compounds. All methanotrophs oxidize methane in a sequential manner, starting with the conversion of methane to methanol, followed by the subsequent oxidation to formaldehyde, formate and ultimately carbon dioxide (Cicerone and Oremland, 1988)

$$CH_4 \rightarrow CH_3OH \rightarrow HCOOH \rightarrow CO_2$$

Methane oxidisers need oxygen for growth and in nature, aerobic methanotrophs will generally be positioned and most active in zones where both methane and oxygen are present. This is usually at the interface between aerobic and anaerobic environments.

Most methanotrophs fix nitrogen, and nitrifiers may also oxidise methane when ammonium is less available. The reason for these double-functions lies probably in the similarity of the two enzymes ammonium monooxygenase and methane monooxygenase. The presence of N in surplus therefore inhibits the CH_4-oxidising activity. Consequently nitrogen fertilisation *in situ* has been shown to significantly decrease net methane uptake by dry soils (Steudler *et al.*, 1989; Mosier *et al.*, 1991). Preliminary data suggest that in tundra environments N fertilisation leads to lower CH_4 emission (K. Nadelhoffer, pers. comm., 1993). If this is due to an effect on CH_4 oxidation the opposite relationship to what was found in the dry soils would be the case. However, it could also be associated with the production of methane. The interaction between nitrogen cycling and methane emission in tundra environments needs further investigation. The possible implications of this question in terms of global change will be discussed more in Chapter 5.

3.4. Stable isotopes and radiocarbon

The discrimination (fractionation) between stable carbon isotopes ^{13}C and ^{12}C by various biogeochemical processes provides a useful tool when analysing carbon cycling. The isotope ratio of a sample is expressed as permil deviation relative to the isotope ratio of an accepted standard (PDB, Craig, 1957).

$$\delta^{13}C = \left[\frac{^{13}C/^{12}C_{sample} - ^{13}C/^{12}C_{std}}{^{13}C/^{12}C_{std}} \right] \times 1000$$

The δD values are calculated similarly in ‰ relative to the SMOW standard (Hagemann *et al.*, 1970).

Photosynthesis discriminates against ^{13}C and most plant tissues (C_3 plants) contain therefore an about 2% (=20‰) lower proportion of $^{13}CO_2$ than the atmosphere. Since atmospheric CO_2 shows an isotopic ratio of -8.0 ‰ versus the PDB standard, most plant tissues show $\delta^{13}C$ values of around -28‰ [i.e. (-8‰) + (-20‰)] (Schlesinger, 1991). The discrimination between $^{12}CO_2$ and $^{13}CO_2$ during photosynthesis is greatest when stomatal conductance is high. The isotopic ratio of plant tissue can therefore be related to the average stomatal conductance during the life history of the plants, providing a long term index of water use efficiency.

Of more direct relevance in the present context is the insight into the methane cycle that the $^{13}C/^{12}C$ ratio may provide when combined with ^{14}C and the hydrogen/deuterium (H/D) of CH_4. These geochemical parameters have revealed at least two different types of methane present in the Earth's crust: methane formed by bacterial production in anaerobic environments ("microbial methane") and methane formed during thermocatalytic reactions ("thermogenic methane") that take place over geological time in association with petroleum formation (Cicerone and Oremland, 1988). In general, microbial gases found in ecosystems are characterised by having methane depleted in ^{13}C, in deuterium, and high in radiocarbon content. The basis for stable isotopic discrimination between microbial and thermogenic gases lies in that methanogenic bacteria enrich for the lighter isotope during methanogenesis. A stable

isotopic scheme has been proposed whereby methane formed through CO_2 reduction can be distinguished from that formed from methyl groups (e.g. acetate) owing to the latter's relative depletion of deuterium (Whiticar *et al.*, 1986). Plots of $^{13}C/^{12}C$ versus D/H ratios (Figure 3.4) are commonly used to distinguish methanogenic pathways and therewith methane sources.

The scheme shown in Figure 3.4 is very general and different processes can produce a more complex picture (see also section 5.3). Microbial methane oxidation, for example, will fractionate in favor of the light isotope, thereby leaving the residual, unoxidised methane enriched in ^{13}C (Coleman *et al.*, 1981). Significant seasonal variability has been observed in the stable carbon and hydrogen ratios of bubble methane reflecting not only varying oxidation rates but also changes in the isotopic signature of substrates like acetate (Cicerone and Oremland, 1988). The fractionation carried out during methane oxidation has been suggested as a tool for attempts to quantify the ratio of methane produced at depth to net emission (Coleman *et al.*, 1981). This could ideally be done by comparing the isotopic signatures of bubble methane, or methane collected from the source region, with methane emitted from the surface after having passed through the aerobic zone. Recent results showing significant isotopic fractionation by plant stems have, however, documented complexities in interpreting such data (Chanton *et al.*, 1992a; Chanton *et al.*, 1992b). Nevertheless, as part of the present study, isotopic data from bubble and diffusion methane were obtained. The data are presented in section 5.3.

Disregarding the problems with interpreting detailed isotopical data, the methane that enters the atmosphere from various sources shows significantly different isotopic signatures (Figure 3.4). The destruction of atmospheric methane by OH (see next section) is selective for the lighter isotopes of carbon and hydrogen, and other atmospheric removal processes may also discriminate. Thus by measuring an isotopic ratio in atmospheric methane along with the isotopic shift in the atmospheric destruction processes, it will be possible to deduce the isotopic ratio of the total source. In general, if two of these quantities are determined, the third is constrained. This

means that it is possible to infer limits on the possible sizes of different methane sources, given that some of them are fairly well known. It is in particular the $\delta^{13}C$ ratios which have been used for constraining the atmospheric methane budget. A detailed discussion of this type of exercise lies beyond the scope of this thesis (see e.g. Quay *et al.*, 1988; Fung *et al.*, 1991; Quay *et al.*, 1991) but the implications they have for tundra methane emission will be touched upon in Chapter 4.

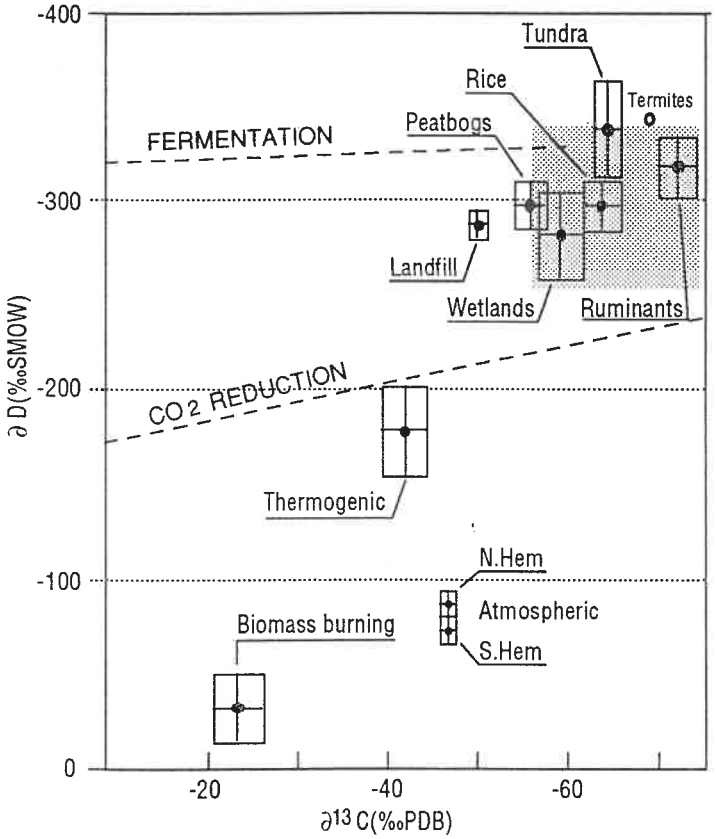


Figure 3.4. Systematics among the stable isotope composition of methane from various sources and from atmospheric methane. Boxes show the mean values for sets of samples for $\delta^{13}C$ and δD in methane as reported by Wahlen (1993). The grey shaded area represents the range of results from analysis of tundra methane as presented in Chapter 5. PDB and SMOW are the respective accepted standards for ^{13}C and D.

Carbon-14 in atmospheric methane can be used to assess the contributions made by fossil (free of ^{14}C) and biogenic (recent ^{14}C) sources in a similar fashion as described above. By means of such analyses the fossil (fuel) contribution to the

atmospheric CH_4 has been estimated to 21±3% of the annual input by Wahlen *et al.* (1989) and 16±12% by Quay *et al.* (1991).

There are, however, complications in using the ^{14}C information from biogenically produced CO_2 and CH_4 to assess age of the source material. This is due to the release of ^{14}C in connection with nuclear bomb testing particularly in the fifties and sixties, and more recently from pressurized light water reactors.

The ^{14}C data in Chapter 5 are reported as pMC (percent modern carbon), the percent with respect to NBS oxalic-acid-based standard activity corrected for decay (Stuiver and Polach, 1977; Levin *et al.*, 1992).

3.5. The role of tundra in atmospheric budgets of CO_2 and CH_4.

3.5.1. Carbon dioxide

According to Adams *et al.* (1990) global peatlands including tundra ecosystems have contributed to remove about 280 Gt of carbon from the atmosphere during the Holocene. If this carbon were to be released at once to the atmosphere, it would increase the present atmospheric concentration of carbon dioxide by about 130 ppm. The uncertainty about the present carbon balance is therefore a critical issue for discussions of future global carbon cycling. There are at present major questions being debated in this regard. The ocean is a significant sink for anthropogenic produced carbon dioxide. Estimates for the size of this sink vary, but it probably amount to about a third of the emissions arising from combustion of fossil fuels. The present atmospheric increase cannot fully account for the remaining emissions and this "missing sink" have given rise to substantial controversy about various estimated figures making up the atmosheric carbon budget. In a recent review of this discussion Siegenthaler and Sarmiento (1993) point to the terrestrial biosphere as the most likely location of this carbon sink. From this viewpoint it seems unlikely that global tundra regions could as yet contribute any large amounts of carbon to the atmosphere. However, as discussed earlier, the potential is present and winter flux measurements (Zimov *et al.*, 1993) in addition to the work by Oechel *et al.* (1993) work mentioned

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above seems to indicate that at least at certain times in particular regions the tundra constitutes a net source of atmospheric carbon dioxide.

3.5.2. Methane

There have been many attempts to investigate the strengths of various methane sources contributing to the atmospheric methane budget. Different approaches include: experimental measurements of methane emission rates in various ecosystems (as in the exercise carried out in Chapter 4) and methane producing processes, investigation of the isotopic composition of methane sources and of atmopheric methane, biostatistical methods, and modelling global atmospheric methane concentration distribution using GCMs. The results of a few recent budget estimates based on various techniques are presented in Table 3.3.

There are basically four major atmospheric sources of methane: natural wetlands and tundra, ruminants, rice production, and fossil methane (natural gas and coal mining). Methane emission from tundra and northern wetlands will be dealt with extensively in the following chapters. Rice fields produce methane in a way similar to natural wetlands, since they are flooded for extensive periods through the growing season and hence provide anaerobic conditions for bacterial methane production. Ruminants also have bacteria as a source of methane, due to enteric fermentation in the rumen. The present increase in atmospheric methane is strongly correlated with the population increase in the third world (IPCC, 1990; IPCC, 1992). This is probably due to the emissions arising from food production in the form of meat and rice cultivation, and underlines the fact that global climatic change as a global environmental issue is strongly linked to the problem of third world population increase.

Fossil methane is being released to the atmosphere in the drilling and exploration of oil and gas. Transmission and distribution of natural gas and coal mining also accounts for significant losses to the atmosphere. Smaller sources include biomass burning, landfills, termites, freshwater and ocean. Methane from clathrate destabilization is a potential enormous source, but its present contribution is probably

small (Kvenvolden, 1988; Kvenvolden and Grantz, 1990; MacDonald, 1990; Kvenvolden *et al.*, 1993). In popular literature this source has often been confused with the emission from continuing microbial production of methane, which is the subject of the present study. This is probably due to large proportions of clathrates being associated with permafrost regions and the possible instability of these following global warming.

As mentioned above, some dry soils act as a sink for atmospheric methane, but the primary removal process is chemically by reaction with OH in the atmosphere (Table 3.3). Atmospheric concentration of methane is therefore highly dependent upon very complex chemical reactions determining the concentration and dynamics of OH. Figure 3.5, referred to as the "flying carpet", shows how the highest atmospheric concentrations of methane are found in the mid- to high northern latitudes seemingly reflecting the large sources (wetlands, tundra, bulk of world population) at these latitudes. However, the patterns shown in Figure 3.5 are also influenced by atmospheric OH chemistry. For example, if only sources were considered, the highest concentrations would be expected in summer where, in particular, wetland emission is peaking. It is, however, in winter and autumn that the atmospheric concentration is highest (Khalil *et al.*, 1993), which probably is due to photochemistry causing OH to be more abundant in summer. Other reasons have also been suggested, and in general the processes governing temporal and spatial variations in the atmospheric methane concentration are still controversial in the literature (Khalil and Rasmussen, 1990a; Fung *et al.*, 1991; Quay *et al.*, 1991; Khalil *et al.*, 1993; Wahlen, 1993). A detailed discussion of this lies beyond the scope of this thesis. The short introduction here serves only to set the scene for the discussion of tundra methane emission in the next chapter.

Table 3.3. Various atmospheric methane budgets as compiled by Wahlen (1993).

	Annual release and range (Tg CH_4/yr)					
	Cicerone and Oremland (1988)		Wahlen *et al.* (1989)	IPCC (1990)		Fung *et al.* (1991)
Source						
Natural wetlands (bogs, swamps, tundra etc.)	115	100-200	147[a]	115	100-200	115
Rice paddies	110	60-170	136	110	25-170	100
Enteric fermentation (animals)	80	65-100	119	80	65-100	80
Fossil methane:			123			
Gas drilling, venting, etc.	45	25-50		45	25-50	40
Coal mining	35	25-45		35	19-50	35
Biomass burning	55	50-100	55	40	20-80	55
Landfills	40	30-70		40	20-70	40
Termites	40	10-100		40	10-100	20
Ocean and fresh waters	15	6-45		15	6-45	10
Hydrate destabilization	5?	0-100		5	0-100	5
Total	540	400-640	580	525	290-965	510
Sink						
Reaction with OH	500	405-595		500	400-600	450
Removal by soils				30	15-45	10
Atmospheric increase		40-46	55	44	40-48	

[a] including landfills

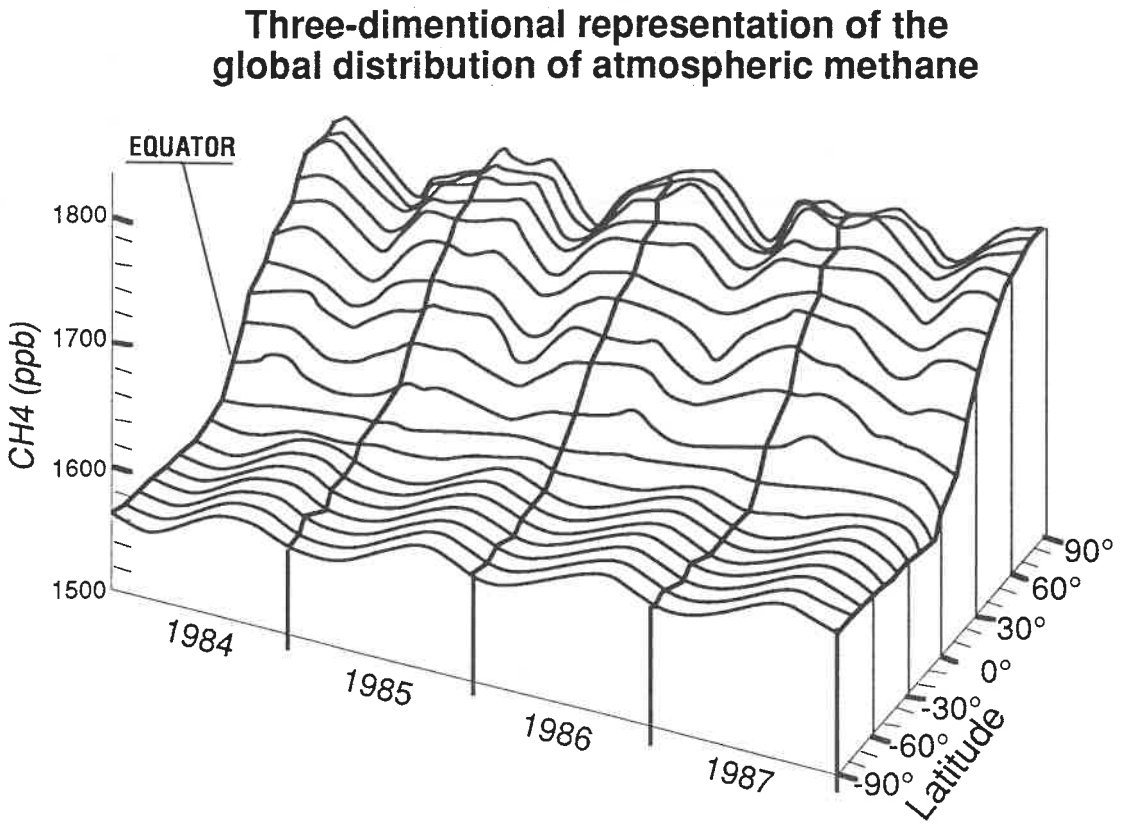

Figure 3.5. Temporal and latitudinal variation of atmospheric methane concentration (from Fung *et al.*, 1991).

3.6. Summary

Following on the description of general "outer" characteristics of tundra presented in the last chapter, in this I have focussed on the processes that leads to trace gas exchange between arctic tundra and the atmosphere.

Annual mean net primary productivity is relatively low in the tundra compared with most other biomes, amounting to about 65 g $C/m^2/yr$. However, this figure covers large differences between tundra subcategories. Shrub tundra environments can fix up to 1000 g $C/m^2/yr$ while dry heath tundra fixes only around 25 g $C/m^2/yr$.

Despite the generally low figures for plant production, tundra environments have accumulated substantial amounts of carbon in the soil. This is due to the cold and wet conditions exerting stronger limitations on total decomposition relative to production. There is a controversy as to the actual amount of carbon stored in the tundra. In

particular, the estimates for soil organic matter in tussock tundra environments vary widely, from 1 to 29 kg C/m^2. Overall areal weighted mean figures for tundra soil carbon varies from 10 to 20 kg C/m^2. The tundra as a whole is probably still accumulating carbon, but it has been shown recently that tussock tundra in Alaska could have shifted from being a sink to a substantial net source of atmospheric CO_2.

Due to the wet conditions on the tundra, anaerobic soil environments are widespread giving rise to substantial methane production. Methane is produced by a group of Archaebacteria which are strictly anaerobic and named "methanogens". Those are highly specialised bacteria that primarily use acetate and hydrogen reduction of CO_2 as substrates for growth. They are dependent upon a network of other microbial interactions for providing them with their substrates. This makes the microbiological constraints on methane production a complex matter. Methanogens are competitively inhibited by the presence of sulfate-reducing bacteria which limits methane production in coastal environments.

The methane which is produced at depth is subject to varying degrees of microbial oxidation. This process has been shown capable of consuming up to 90% of the potential emission and makes microbial oxidation highly important in controlling net emission of methane from tundra environments to the atmosphere. The oxidation is influenced by the available nitrogen in the soil which links controls on methane emission to those on nutrient cycling in the soil.

Isotopical data give considerable insight into varies aspects of carbon biogeochemistry and, for example, are widely used as constraint on atmospheric methane budgets. Various methane sources have distinct isotopical "signatures". In general biogenic produced methane is characterised by being depleted in both $\delta^{13}C$ and δD. Microbial methane oxidation can be identified and potentially quantified by isotopical analyses since this process fractionates in favor of the light C isotopes, thereby leaving the residual, unoxidised methane enriched in ^{13}C.

Chapter 4

Methane emission from tundra ecosystems

4.1. Introduction

The question I address in this chapter is fundamental but not very well understood: what is the present scale of methane emission from tundra environments? A major part of the discussion will be based on data from my field work in Alaska. However, I rely also on comparison with other flux studies, in particular the work by Whalen and Reeburgh (1988; 1992), in order to validate my own flux estimates. Various ways of extrapolating the data to regional and global flux estimates will be discussed.

4.2. Toolik Lake field study

Methane flux was measured in summer 1991 and 1992 at permanent tundra sites near the University of Alaska Field Station at Toolik Lake (68°38'N, 149°38'W), 650 kilometres north of Fairbanks on the North Slope of Alaska (Figure 4.1). There were several reasons for choosing the Toolik Lake area as subject for intense field study:

1) Toolik Lake is situated in the "typical tundra" zone (see Chapter 2) in an area with extensive tussock tundra. Wet and low shrub tundra environments are also present. The area is dominated by rolling hills, and local topographical transects provided opportunity for studying natural moisture gradients.

2) There is relative easy access to Toolik Lake from Fairbanks by the Trans-Alaska Pipeline Haul Road. The area is unique to the world in the sense that it is possible to drive to true arctic terrestrial field sites (Figure 4.1).

3) The field station provides the necessary logistical facilities for operating a gas chromatograph.

4) Toolik Lake represents the tundra biome in a NSF funded Long Term Ecological Research (LTER) programme studying major biomes of the U.S. This programme is producing basic ecological and meteorological monitoring data from the area which

have been very useful for the present study. For example the modelling effort described below in Chapter 6 could not have been carried out the way it was without the weather data provided by the LTER programme.

5) Collaboration with Professor W.S. Reeburgh and Dr S.C. Whalen in Fairbanks on methodology and use of equipment was very beneficial and made a direct comparison of sub-arctic and true arctic tundra methane flux possible.

Figure 4.1. Alaska showing the Dalton Highway and the location of Toolik Lake (68°38'N, 149°38'W).

All sites are in the continuous permafrost zone. Measurements were made at 22 stations, on 23 days through the 1991 thaw-season and on eight days during August 1992. Of these stations, 16 were established by Whalen and Reeburgh in 1987; the remainder were established for this study. Four extra stations were established in 1992 on tussocks (T sites) with approximately 50 meter intervals along a natural moisture gradient up-slope from the Kuparuk River (Figure 4.2 and 4.3).

The stations are grouped in six floristically different units representative of arctic tundra:

1) *Eriophorum* tussocks (T sites).
2) Waterlogged intertussock areas without vegetation ("black holes", BH sites, Figure 4.4).
3) Mosses invaded to various extent by vascular plants (M sites).
4) *Carex* at a pond margin mixed with *Eriophorum* (C sites) .
5) "Depressed" waterlogged areas with *Eriophorum* and *Carex* (D sites).
6) "Elevated" areas surrounding the depressions with a variety of plants such as mosses and vascular plants including small shrubs (E sites).

C and D sites are typical for wet tundra while BH (although waterlogged), M and T sites all form building stones of tussock tundra environments. E sites resemble dwarf shrub/sedge tundra. All floristic units except D and E have directly comparable sub-arctic counterparts in studies by Whalen and Reeburgh (1988; 1992) forming basis of a comparison between sub-arctic and arctic tundra fluxes.

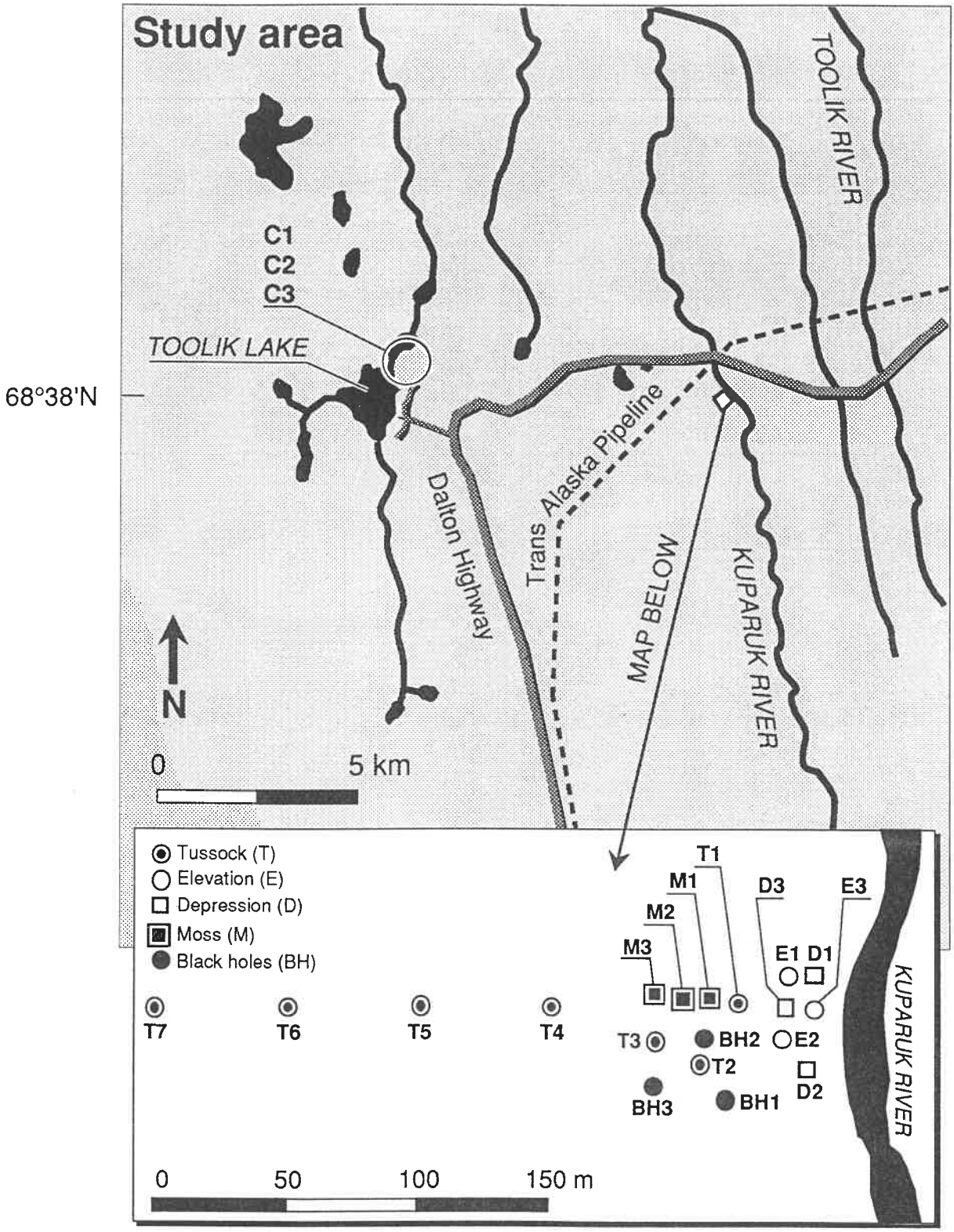

Figure 4.2. Map showing study site.

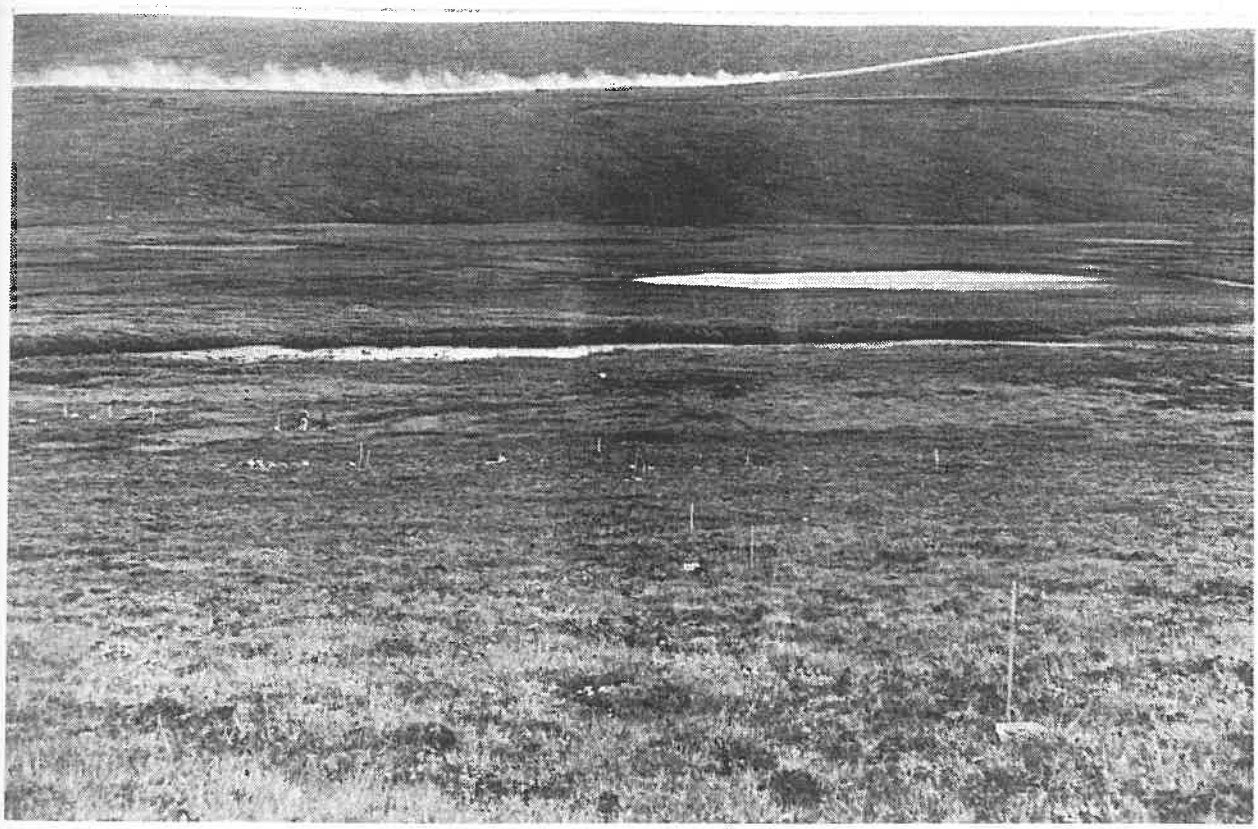

Figure 4.3. Photo showing main study site. The dust in the background is from a truck heading north on the Dalton Highway.

There are unfortunatly very limited historical weather data available from the Toolik Lake area. Data-logging of climate was not initiated before the LTER programme and the long term record at Toolik Lake therefore dates back only to 1988 (LTER, 1992). At Imnavait Creek watershed 11 kilometres east of Toolik Lake, the Water Research Center at University of Alaska has been monitoring weather since the early eighties. Here their data were used complementory to the LTER data. There are differences, however, in weather at Toolik Lake and Imnavait Creek.

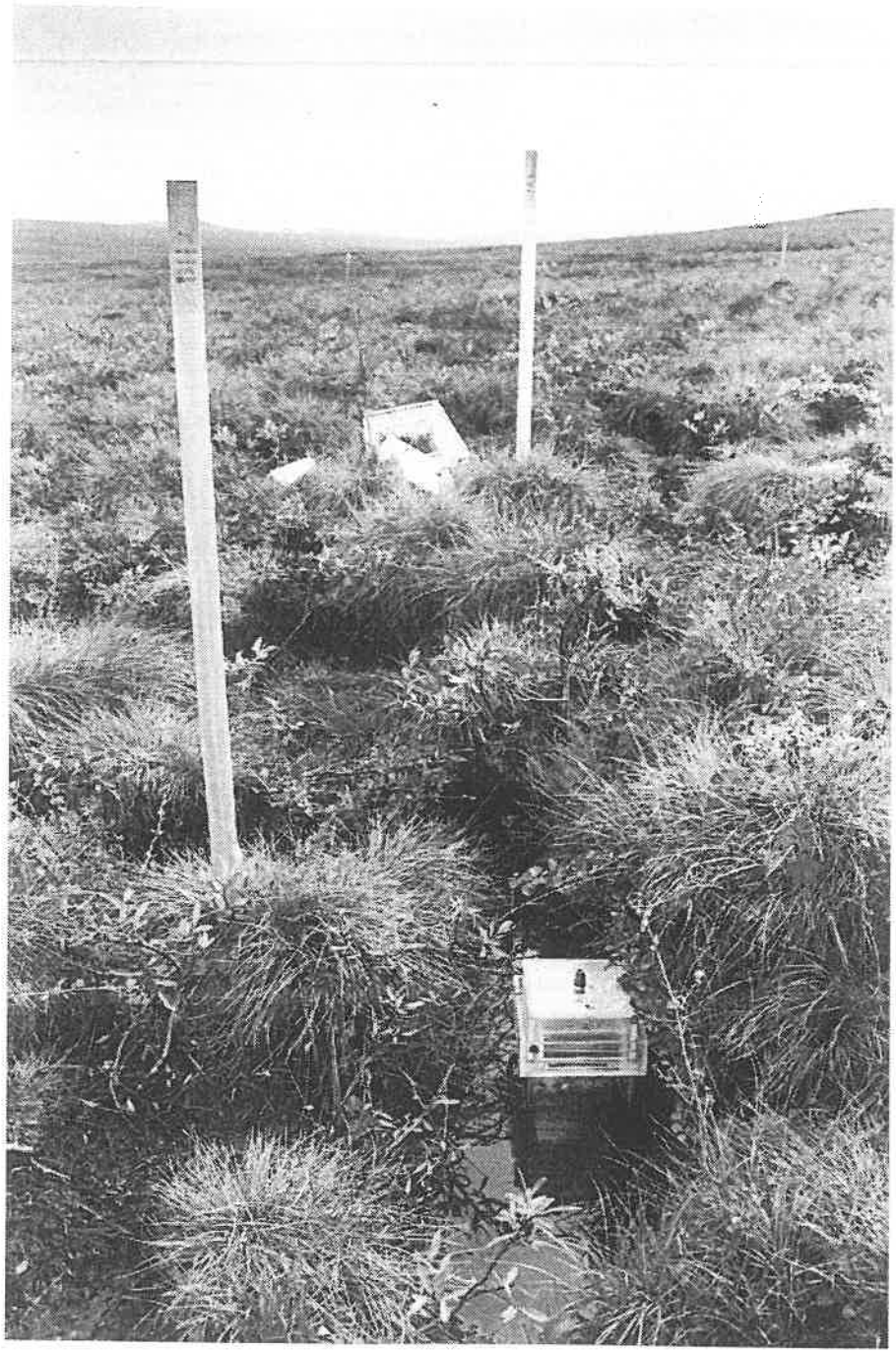

Figure 4.4. Chamber installed at BH site surrounded by tussocks.

The mean air temperature at Toolik Lake in summer (June-August) 1991 was 7.9°C considerably lower than the four-year average of 9.8°C, and closer to the long term average of 8.4°C at Imnavait Creek. It seems, though, that 1991 was a relatively cold year. In particular August stood out with a mean temperature of 5.6°C and minimum temperatures as low as - 9.9°C.

Considering Toolik Lake only, summer in 1991 precipitation was quite normal, with 138 mm compared with a mean of 126 mm. However, it is low compared to Imnavait Creek where the long term mean of accumulated precipitation June through August is 202 mm.

The data-logger at Toolik Lake was not functioning consistently in June and July 1992, which makes all summer comparisons impossible (LTER, 1992). The August 1992 mean temperature of 7.3°C was slightly lower than the five-year August average of 8.6°C. In terms of precipitation August 1992 at Toolik Lake stood out with a massive rainfall in late August. Accumulated precipitation in August alone amounted to 118 mm. However, the bulk of this rain fell during two days immediately after my field work had terminated on 24 August and the possible effects it would have on trace gas exchange were therefore not detected.

The methods used for gas sampling and analysis and for measuring environmental variables are described in Appendix 1.

4.2.1. Emission from floristic units

The mean, maximum and median emissions as measured at all stations in 1991 and 1992 are presented in Table 4.1. The emissions found in earlier, more sporadic, measurements at the same sites (Whalen & Reeburgh, unpublished) were all within the ranges found in this study.

Flux measurements in 1992 were limited timewise. Obviously, there are significant annual variations in flux and, as discussed in Chapter 2, the progressing season on the tundra is best described as a function of degree days. Therefore, in order to compare 1991 and 1992 equivalent degree days above 0°C were used. Intensive flux

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measurements were carried out between 1 and 11 August 1992 (degree days 596-699). This matches the period between 25 July and 9 August 1991 (degree days 591-699), where an overall mean flux from all sites shows 43% higher flux in 1992 compared with 1991. Chapter 5 will deal in detail with controls on the scale of emission, and possible reasons for higher flux in 1992.

Table 4.1. Summary of CH_4 flux observed during the 1991 thaw season and in August 1992 on the North Slope of Alaska. Stdev is the standard deviation of mean.

		CH_4 flux mg/m²/day			
	n	Mean	Stdev	Max	Median
BH1	28	0.45	0.53	2.25	0.3
BH2	28	2.4	2.78	11.45	1.58
BH3	28	0.16	0.24	0.83	0
M1	28	12.85	5.09	23.26	13.59
M2	28	2.5	2.32	8.53	2.18
M3	27	18.28	7.65	36.57	19.61
T1	28	95.25	74.03	252.41	82.72
T2	28	35.15	18.9	65.63	36.46
T3	28	32.09	26.71	114.73	25.4
T4[a]	5	5.45	0.71	6.5	5.47
T5[a]	5	0.37	0.52	1.03	0
T6[a]	5	0.19	0.42	0.95	0
T7[a]	5	-0.32	0.32	0	-0.4
C1	27	36.63	39.76	176.19	20.95
C2	27	373.28	567.83	2228.17	151.58
C3	27	39.39	23.73	108.34	34.53
D1	28	116.62	36.57	205.51	124.07
D2	28	35.39	24.81	116.8	29.38
D3[b]	21	50.06	28.83	109.56	53.04
E1	28	34.44	25.94	104.12	27.24
E2	28	47.51	29.94	132.67	42.34
E3[b]	19	10.79	4.34	16.53	11.96

[a] measured only in 1992

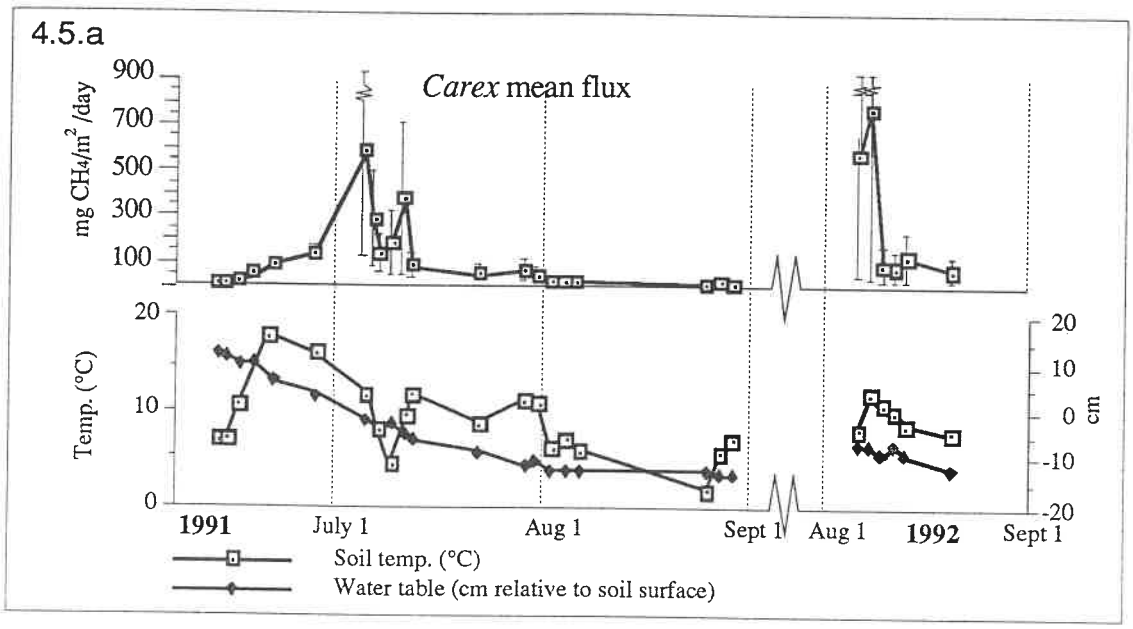
[b] measured only in 1991

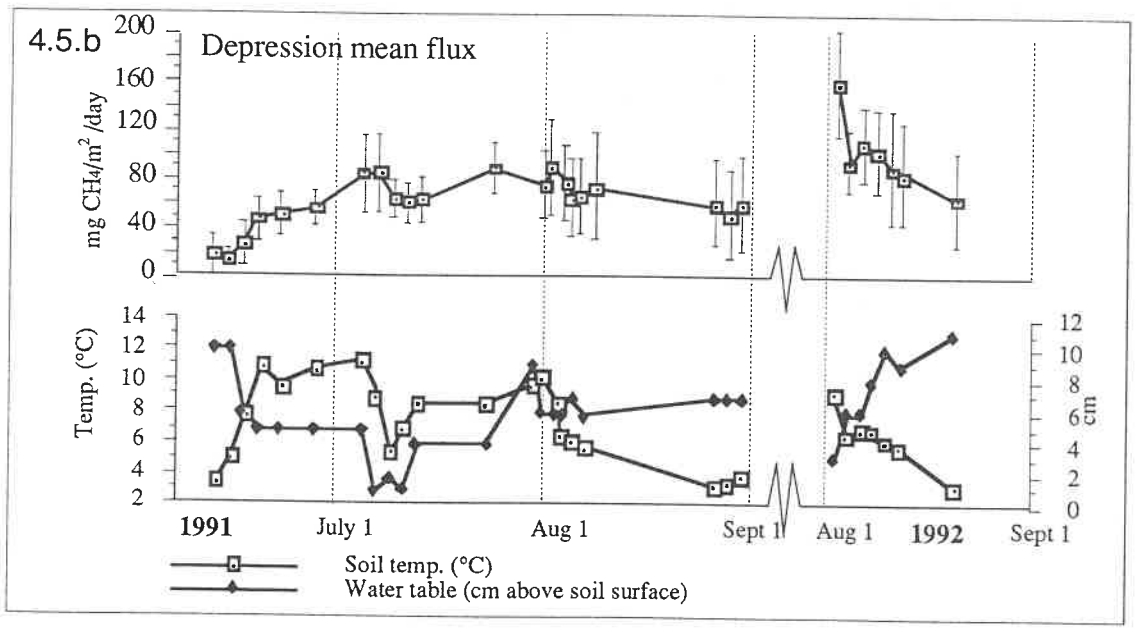
Fluxes at three sites were used in 1991 to calculate a seasonal integrated flux from each floristic unit. The thaw-season was estimated to last for 100 days, based on weather information from the region supplied by the Water Research Center, University of Alaska, and unpublished LTER weather data from Toolik Lake (LTER, 1992). In order to encompass the thaw season, one week was added to the integrated curve before the first sampling date, and two weeks after the last, to allow emissions to approach zero. To calculate the annual emission, a factor representing the assumed winter flux based on observations by Whalen and Reeburgh (1988) was multiplied by the seasonal integrated flux (winter emission as percentage of annual flux: *Carex* 9%, tussocks 5%, black holes 4%, mosses 10%, depressions 9% estimated, and elevations 5% estimated).

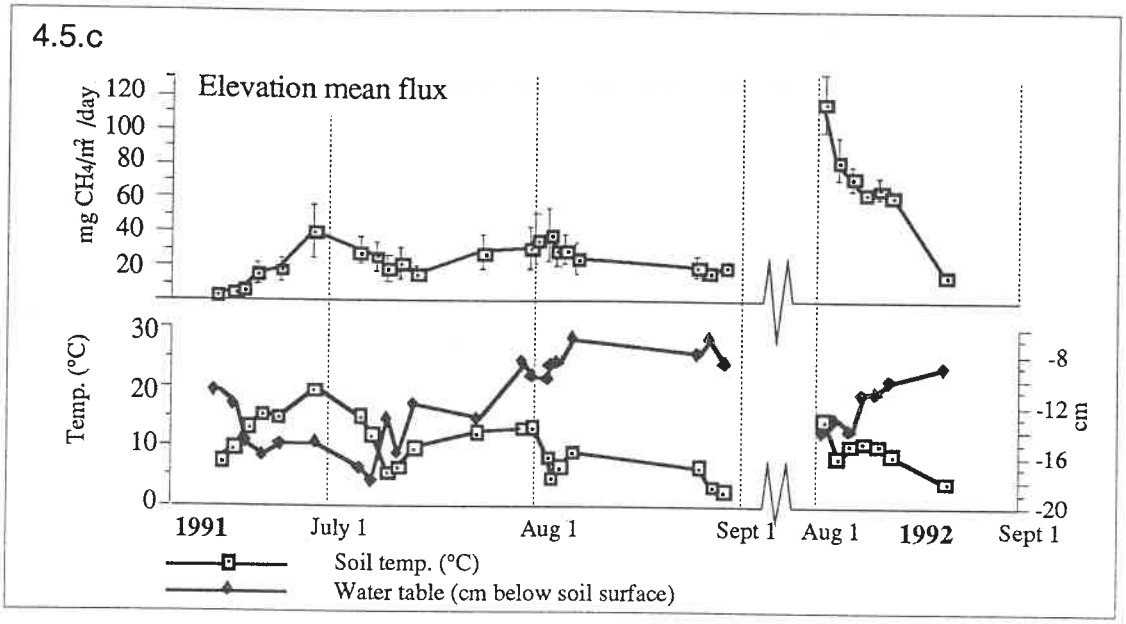
As observed in most other field studies of CH_4 emission, the sites showed a substantial natural variation. The standard deviation (Table 4.1 and 4.2) and standard error (Figure 4.5) were particularly high at the *Carex* sites due to episodic events of very high emission. The daily mean and seasonal integrated fluxes as measured at the different units are presented in Table 4.2. Also in Table 4.2 are shown the corresponding ranges of flux found in Whalen and Reeburgh's four year time-series of measurements at corresponding sub-arctic sites in the Fairbanks area.

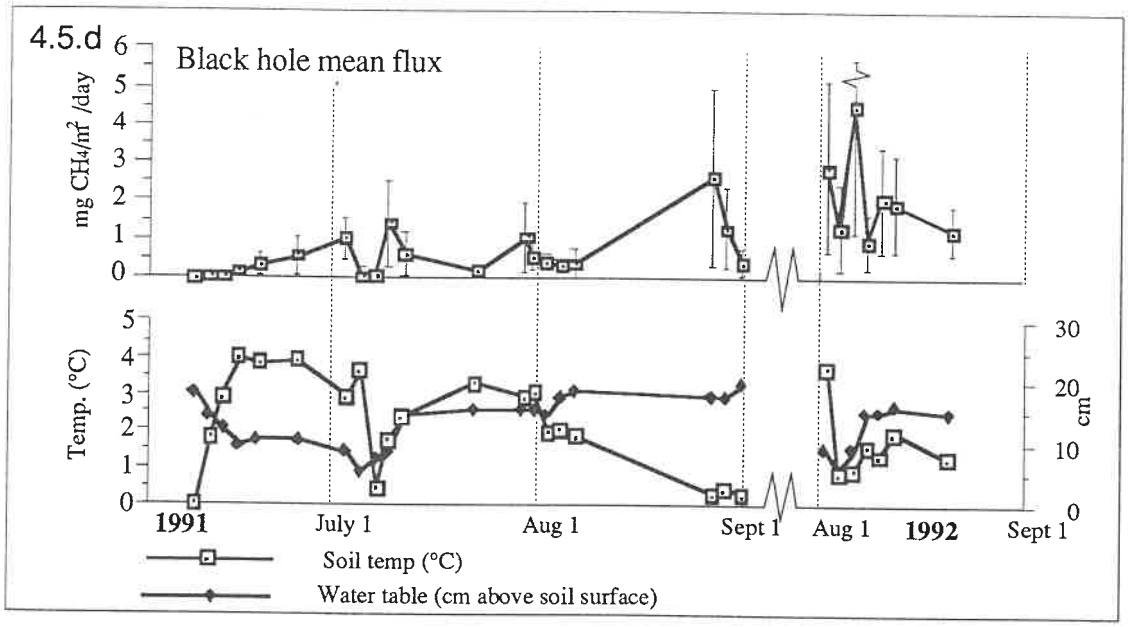
All fluxes are somewhat higher in the sub-arctic than at the arctic sites but only the intertussock fluxes show a significant difference. As mentioned above, the large range at the *Carex* sites on the North Slope is due to episodic events of extreme emission. The similarly large range in the Fairbanks *Carex* data, however, is not due as much to interseasonal variations but rather that one whole year (1990) stood out with integrated emission more than ten-fold above other years. This could be due to a change of *Carex* sites carried out in 1990, but Whalen and Reeburgh claim similar sites were found and that spatial differences cannot fully explain the extreme emission. Rather, according to Whalen and Reeburgh, the difference were due to record rainfall, hence climatically controlled (Whalen and Reeburgh, 1992)

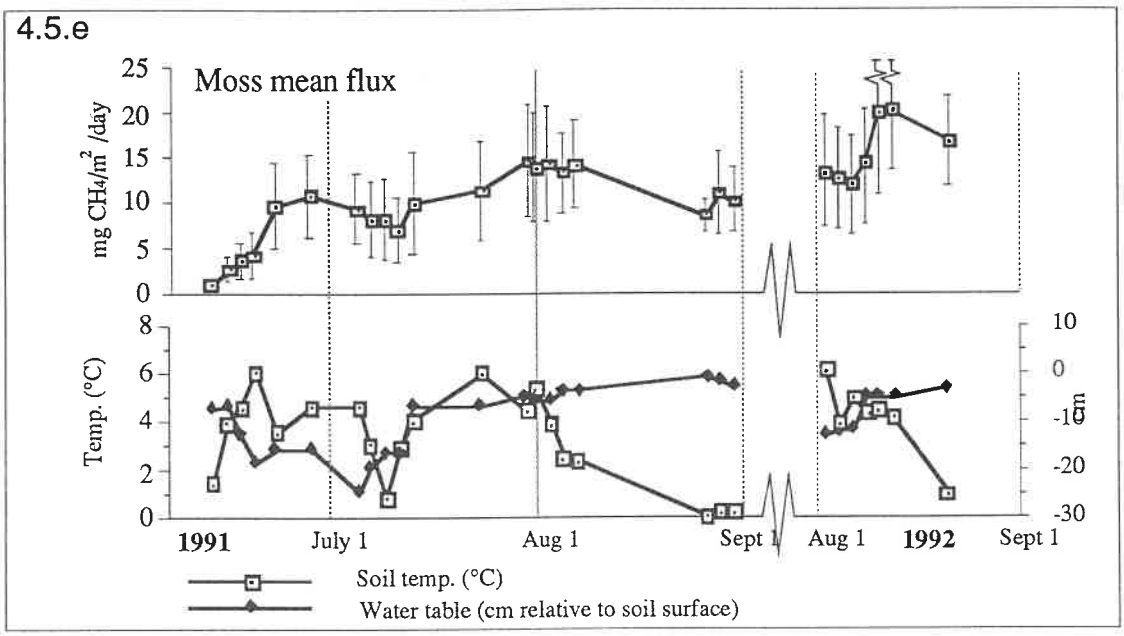
Figure 4.5. Mean CH_4 flux, soil temperature (mean of top 13 cm) and water table height at six different floristic units (a: *Carex*; b: depression; c: elevation; d: black hole; e: moss; f: tussock) through the 1991 and 1992 thaw seasons near Toolik Lake on the North Slope of Alaska. Error bars indicate standard error of means (n = 3). Bars are absent where standard error is smaller than symbol.

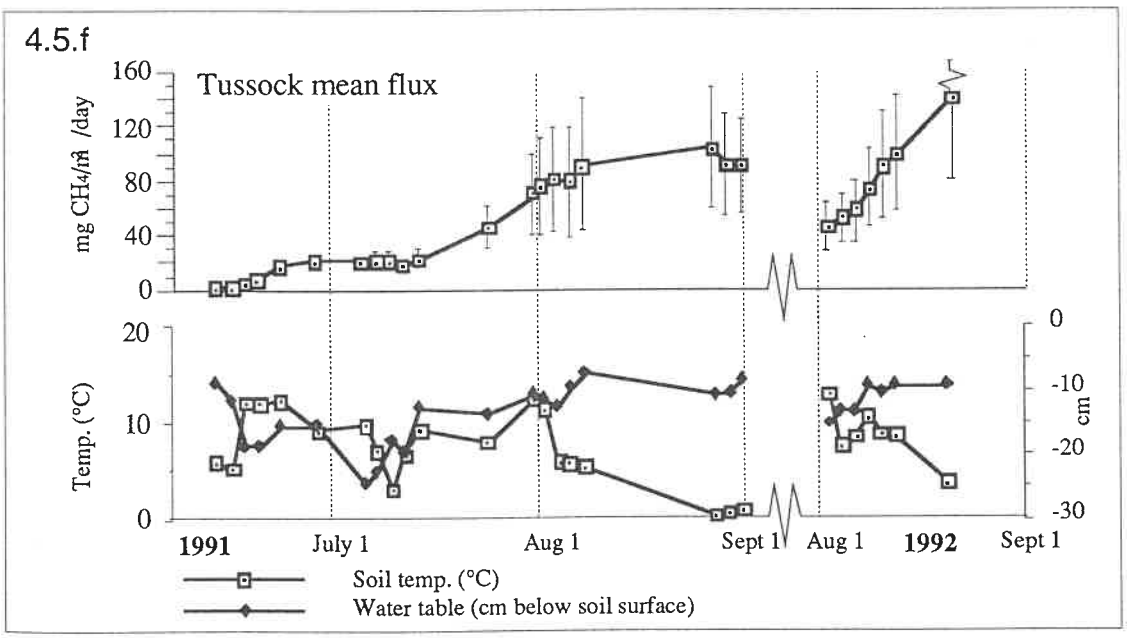


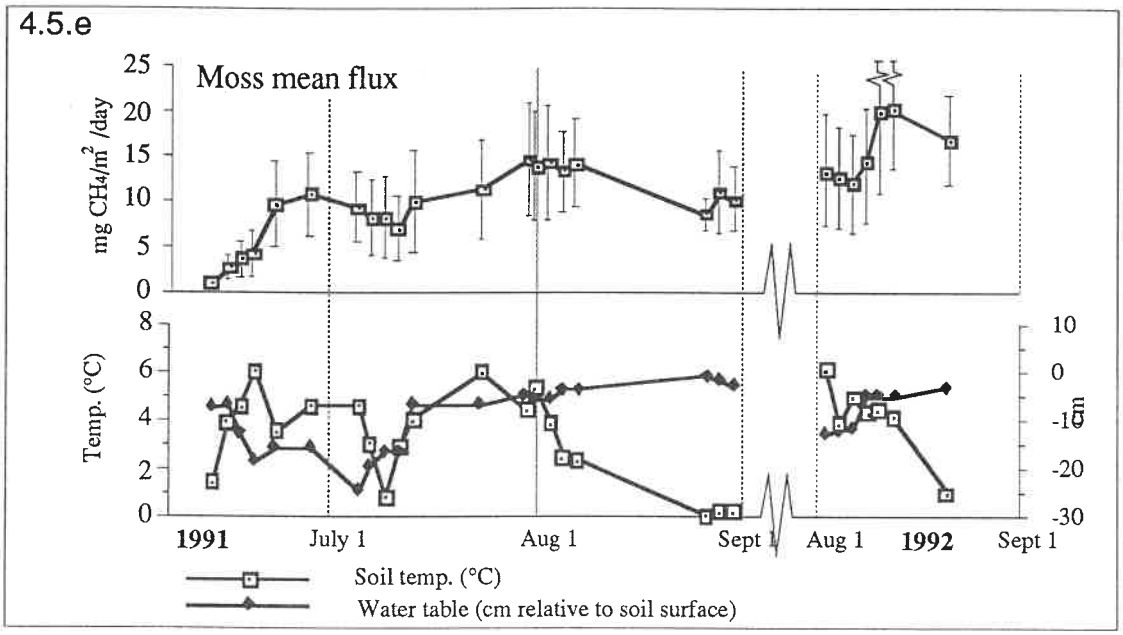



4.5.c
Elevation mean flux
mg CH4/m2 /day
120
100
80
60
40
20
Temp. (°C)
30
20
10
0
-8
-12
-16
-20
cm
1991
July 1
Aug 1
Sept 1
Aug 1
1992
Sept 1
Soil temp. (°C)
Water table (cm below soil surface)


4.5.d
Black hole mean flux
mg CH4/m2 /day
6
5
4
3
2
1
0
Temp. (°C)
5
4
3
2
1
0
30
20
10
0
cm
1991
July 1
Aug 1
Sept 1
Aug 1
1992
Sept 1
Soil temp (°C)
Water table (cm above soil surface)


4.5.e
Moss mean flux
mg CH4/m2 /day
25
20
15
10
5
0
Temp. (°C)
8
6
4
2
0
10
0
-10
-20
-30
cm
1991
July 1
Aug 1
Sept 1
Aug 1
1992
Sept 1
Soil temp. (°C)
Water table (cm relative to soil surface)


4.5.f
Tussock mean flux
mg CH4/m2 /day
160
120
80
40
0
Temp. (°C)
20
10
0
0
-10
-20
-30
cm
1991
July 1
Aug 1
Sept 1
Aug 1
1992
Sept 1
Soil temp. (°C)
Water table (cm below soil surface)


4.5.e
Moss mean flux
mg CH4/m2 /day
25
20
15
10
5
0
Temp. (°C)
8
6
4
2
0
10
0
-10
-20
-30
cm
1991
July 1
Aug 1
Sept 1
Aug 1
1992
Sept 1
Soil temp. (°C)
Water table (cm relative to soil surface)

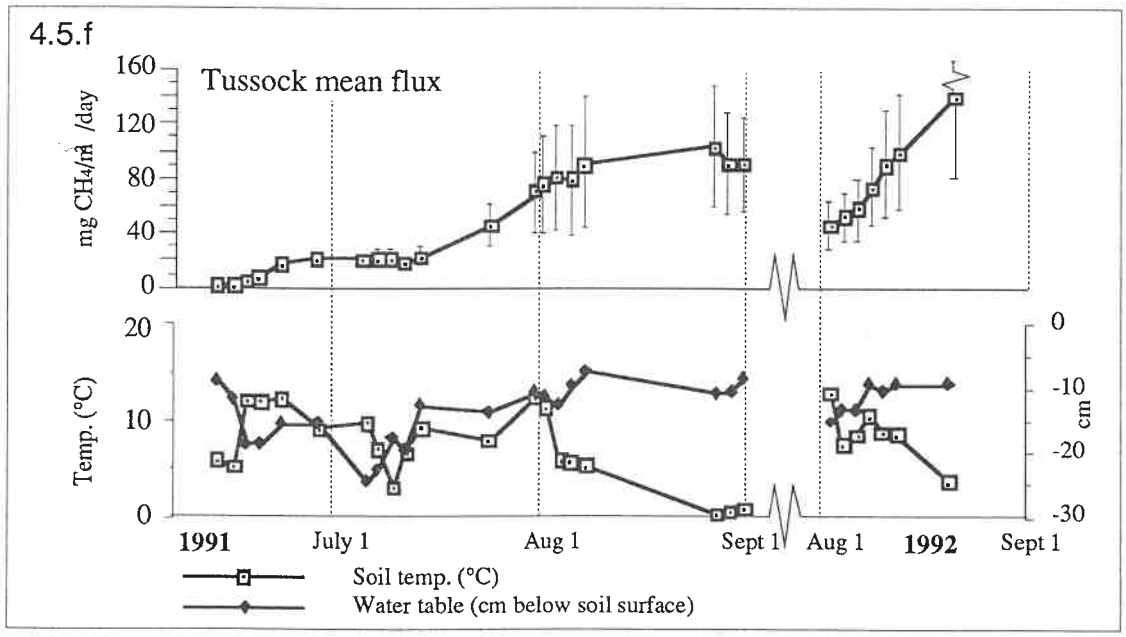

4.5.f
Tussock mean flux
mg CH4/m2 /day
160
120
80
40
0
Temp. (°C)
20
10
0
0
-10
-20
-30
cm
1991
July 1
Aug 1
Sept 1
Aug 1
1992
Sept 1
Soil temp. (°C)
Water table (cm below soil surface)

Table 4.2. The daily mean and integrated annual CH_4 flux from sites on the North Slope in 1991 and in the Fairbanks area of Alaska in 1987-90. The daily mean is of thaw season flux in three chambers ± standard error (mg m^{-2} day^{-1}). Net annual CH_4 flux (g m^{-2} yr^{-1} ± standard deviation) from the North Slope is integrated under a mean flux curve by the trapezoidal rule. The corresponding estimates for the individual years in Fairbanks were calculated similarly but the figures in the last column are means (± standard error) of the figures for each of the four years. The net flux from the North Slope is based on a total number of measurements averaging 60 for each unit, multiplied by a factor representing the winter flux (see text).

Site type	Daily mean CH_4 flux	Fairbanks equivalent	Integrated net CH_4 flux	Fairbanks equivalent
Tussock	42.8±16.3	59.0±12.0	4.73±1.2	10.3±1.4
Intertussock	0.6±0.4	11.6±3.8	0.067±0.04	1.9±0.8
Carex	112.4±71.8	127.9±60.6	8.83±3.4	17.6±14.4
Moss	9.2±3.8	11.0±6.2	1.24±0.14	2.5±1.2
Depressions	61.6±24.8	--	6.0±1.1	--
Elevations	22.7±7.2	--	2.1±0.27	--

Interseasonal episodic events were, in the present study, measured repeatedly in 1991 and 1992 at one site (C2). In the first two weeks of July 1991, this site emitted CH_4 at a rate more than seven times higher (1.5 g CH_4/m^2/day) than the normal range for *Carex* (<200 mg CH_4/m^2/day, Figure 4.6). On 5 August, 1992, C2 reached a peak emission of 2.2 g CH_4/m^2/day and the mean *Carex* emission a peak of 767 mg CH_4/m^2/day (Figure 4.5.a). The former is the highest tundra methane emission ever reported (Christensen, 1993). C2 also showed the highest emission in the limited early measurements (Whalen & Reeburgh, unpublished).

Occasional very high methane fluxes have been reported by many investigators since first observed by Clymo and Reddaway (1971), but explanations for the events have been rare. The episodic events at C2 seem not to be related to any similar changes in measured environmental factors. They are unlikely to be due to physical releases of large methane bubbles since such releases would have disturbed the linearity of concentration change in the sampling chamber with time. Also the events lasted

consistently for several days (Figure 4.6), which is unlikely for a sudden physical release of large methane bubbles. However, although it was not observed during the measurements, a steady stream of smaller bubbles cannot be ruled out as cause for the events, given the sampling system used. Windsor *et al.* (1992) show how such episodic emission could be associated with a reduction in overburden pressure followed by a lowered water table. This is probably not the reason for the events reported here since no similar fluxes were observed at C1 and C3, which experienced similar fluctuations in water table as C2. The events could also be associated with "hot spots" for microbial activity, which have been reported to occur due to the presence of particulate organic carbon in the soil (Parkin, 1987). In any case the events reported here illustrate the importance of detailed time series of flux data when the aim is to extrapolate and estimate annual emissions. The frequency and scale of these events would have a major impact on the global estimates if neglected or not measured. Windsor *et al.* (1992) report that seasonal estimates become 7-22% lower when measured episodic events are excluded from calculations. The global estimate of this study would be 21-25% lower if the episodic event were excluded.

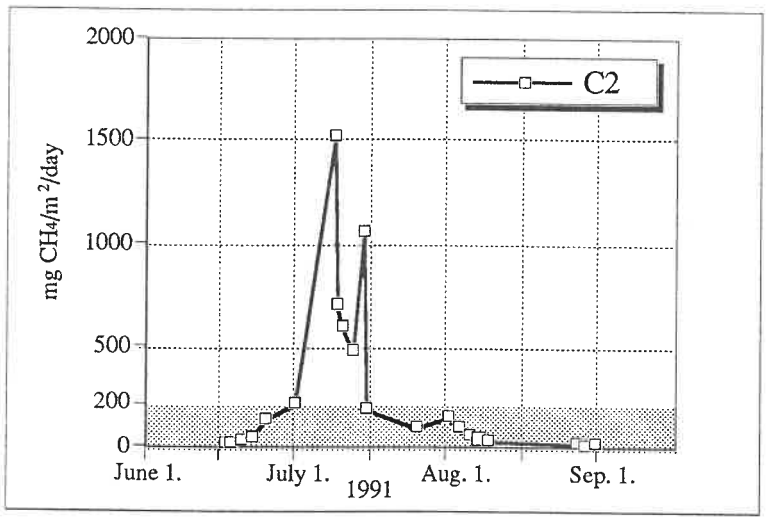


Figure 4.6. Episodic event of CH_4 emission at a *Carex* site (C2). Previous measurements at all *Carex* sites and the 1991-92 data for C1 and C3 all have emission rates in the grey area (below 200 mg CH_4 m^{-2} day^{-1}).

Integrated tussock flux on the North Slope and in the Fairbanks area (T sites) are both somewhat higher than measured by other studies (Bartlett *et al.*, 1992; Morrissey and Livingston, 1992; Torn and Chapin, 1993). The 1992 study of additional T sites on a transect with a natural gradient in water table position showed rapidly decreasing CH_4 flux when moving up-slope into dryer tundra areas (T4-7 in Table 4.1, see section 5.2). This indicates that the 1991 data from moist T sites, used for the above integration, might overestimate average tussock flux, since the 1991 T sites (T1-3) are the downslope part of the transect. The top chamber on the transect also showed consumption of atmospheric CH_4 by tundra soils, but not enough data were obtained for determining the significance of this in terms of global tundra methane emission estimates. However, Bartlett *et al.* (1992) and data from dryer sites in a transect study from Prudhoe Bay to the Arctic Circle in 1992 (Whalen & Reeburgh, unpublished) also showed negative net CH_4 fluxes, so it seems that it may be necessary to account for a small dry tussock sink in extrapolations of tussock flux.

BH sites show the lowest emission observed anywhere for this unit (0.6±0.4 mg $CH_4/m^2/day$) and have significantly lower emission on the North Slope when compared to Fairbanks (11.6 mg $CH_4/m^2/day$). Although waterlogged, the constantly low temperatures (due to shading by the tussocks) and shallow organic layer in the soil (Table 5.1) might explain the low flux. BH emission could be considered below the "noise" level.

Emission at D sites (61.6±24.8 mg $CH_4/m^2/day$) are in the general range of *Carex* and are probably representative of the emission from wet meadow tundra environments. E sites have microtopographical and floristic characteristics in common with both tussocks and mosses and could, as the emission (22.7±7.2 mg $CH_4/m^2/day$) also indicates, be considered a combination between the two.

In general vascular plant communities (C, T, D and E) have the highest emission. This is in agreement with most other flux studies. The flux:plant relationships have been quantified as functions of plant biomass (Morrissey and Livingston, 1992;

Whiting and Chanton, 1992) and also of the number of vascular plant tillers present in the sampling chamber (see section 5.4.3).

Table 4.3. Range of methane emission from major tundra subdivisions as measured for tussock, wet, and low shrub tundra and estimated from various sources for polar desert, semidesert and tall shrub tundra. The area covered by each subdivision is in Table 3.1

	mg $CH_4/m^2/yr$
Desert	-2 - 2
Semidesert	-2 - 10
Wet sedge	20 - 2200
Tussock	-2 - 252
Low shrub	-2 - 50
Tall shrub	-2 - 5

4.2.2. Flux from tundra subdivisions

The ranges of methane emission from major tundra subdivisions are presented in Table 4.3. Looking more closely at the two major tundra types on North Slope tussock and wet tundra, the 1991 data from Toolik Lake are used for comparisons due to the limited data set from 1992. On the basis of daily mean flux (Table 4.2) and percentage coverage of each unit (as used for 1991-92 data in Table 4.4 and 1991 global estimate in Table 4.5) the 1991 thaw season mean of wet meadow tundra flux for the area surveyed is calculated to be 97±6 mg $CH_4/m^2/day$ (range due to uncertainty in percentage cover estimates). This is comparable with an overall estimate (110±6 mg $CH_4/m^2/day$) based on a four-year average (Whalen and Reeburgh, 1992) from the sub-arctic and the transect-based estimate of 90 mg $CH_4/m^2/day$ obtained in 1987 partly in the same region as this study (Whalen and Reeburgh, 1990b). The result presented here is higher than Morrissey and Livingston's (1992) recent estimate of 64 mg $CH_4/m^2/day$ for 1987 wet meadow flux in the same region as the present study.

Thaw season tussock tundra flux is calculated as 25±8 mg CH_4/m^2/day based on 1991 data (site T1-3 only). It is in line with Whalen and Reeburgh's transect result of 31 mg CH_4/m^2/day (Whalen and Reeburgh, 1990b) and within range of the multi-year Fairbanks average of 35±10 mg CH_4/m^2/day (Whalen and Reeburgh, 1992). Bartlett *et al.* (1992) and Morrissey and Livingston (1992) found substantially lower tussock tundra emissions (around 3 mg CH_4/m^2/day). The tussock sites used for the integration in this study and by Whalen and Reeburgh (1992) are clearly part of the tussock tundra environment, but there is very little overlap with the tussock tundra flux range observed by Bartlett *et al.* (1992) and Morrissey and Livingston (1992). This might indicate that these two studies could have a bias towards dry tussock tundra in a similar fashion to the above mentioned wet tussock bias in the present study. The results show, in any case, the variety of environments covered by the term tussock tundra.

4.3. Northern wetland flux

Most extrapolations of northern wetland/tundra emission include large non-permafrost areas in boreal and temperate regions. These areas are more heavily surveyed than more northern regions and true tundra in particular. I carried out an experiment with soil cores from a temperate bog environment in Bøllemosen (55°50' N, 12°36' E) 15 kilometres north of Copenhagen, Denmark. The experiment was designed to test the temperature dependency of saturated soils (see section 5.3.) and the prime objective was thereforee not to measure the scale of emission *in situ*. However, the mean emission from the soil cores when kept between 5 and 15°C is thought reasonably representative of emission from the site. The emission from the open non-forested part of the bog was 17.6±3.9 mg CH_4/m^2/day (n=44) and from the forested bog margin 3.2±1.4 mg CH_4/m^2/day (n=40). The same significant difference in flux from non-forested and forested bog environments have been found in many studies (Crill *et al.*, 1988; Roulet *et al.*, 1992; Roulet *et al.*, 1993). However, the scale of emission varies greatly, particularly between studies of low boreal wetlands. Open bogs in Minnesota were shown by Crill *et al.* (1988) to emit an average of 294±30 mg CH_4/m^2/day while

Roulet *et al.* (1992b) in central Ontario found bog emissions in the range of 6 - 21 mg $CH_4/m^2/day$, which is in agreement with the emission I found in Denmark. Thicket swamps and beaver ponds had the highest emission in the boreal wetlands surveyed by Roulet *et al.* (1992b) (0.1 - 88 mg $CH_4/m^2/day$ and 30 - 90 mg $CH_4/m^2/day$, respectively) but in general very low emission was found.

Forested parts of the Minnesota bogs had emissions of 77±21 mg $CH_4/m^2/day$ (Crill *et al.*, 1988) whereas the corresponding figure was much lower in central Ontario (< 8 mg $CH_4/m^2/day$ (Roulet *et al.*, 1992)). Again the latter area seems to have fluxes similar to the Danish wetland I surveyed.

From a major study of high boreal and sub-arctic wetlands in Canada Roulet *et al.* (1993). report relatively low fluxes. Using Landsat Thematic Mapper images to yield a habitat-weighted emission based on ground based flux studies of 16 habitats in a high boreal and five habitats in a sub-arctic wetland, they found averages of 16 and 44 mg $CH_4/m^2/day$ respectively.

In late July 1993 I carried out a preliminary study of CO_2 and CH_4 flux in a sub-arctic bog near Abisko (Stordalen), Northern Sweden. The bog was the same as that extensively surveyed by Svensson and others (Svensson, 1976; Svensson, 1980; Svensson and Roswall, 1984) in connection with the International Biological Programme (IBP) in the seventies. My data from 1993 show rather low emissions compared to Svensson's earlier work; I found a mean emission of 9.2 $mg/m^2/d$ (range 0.7-60.6) at wet sites (>1000 %dw) and 0.5 $mg/m^2/d$ (range -0.4-2.1) at hummock sites (< 1000 %dw) (see section 5.5.2). These figures are generally lower than the range of 8.6-950 $mg/m^2/d$ reported by Svensson and Roswall (1984)

In summary boreal and sub-arctic wetlands show highly varied emission ranging from 9 to 300 $mg/m^2/day$ for open bogs and 3 to 80 $mg/m^2/day$ for forested bogs. For more extensive reviews of the literature on methane emission from northern wetlands see Bartlett and Harriss (1993) and Harriss *et al.* (1993).

4.4. Extrapolations

Any extrapolation of methane flux based on a limited number of measurements in a geographically restricted area involves substantial uncertainty. However, this type of data forms important sources of validation for the atmospheric budgets proposed by various methods, mentioned in section 3.5. In the following, three different approaches will be described. They all contain substantial uncertainties, particularly in the global extrapolations. The main purpose of the discussion is to provide some insight into the problems associated with any extrapolation of this kind.

4.4.1. Alaskan tundra methane flux

The most extensively surveyed tundra types in this study are tussock and wet sedge tundra, and the data are from Alaska. The most immediate extrapolation is therefore to flux estimates from these two tundra types restricted to Alaska.

Using mean emission from 1991-92 and data on percentage coverage of the individual floristic units, tussock tundra emission is calculated at 15-29 mg/m^2/d (Table 4.4). This leads to an Alaskan tussock tundra methane emission estimate of 0.2-0.4 Tg/yr. Similarly emission from wet sedge tundra amounts to 88-100 mg/m^2/d corresponding to a total of 0.9-1.0 Tg/yr. In both cases the daily mean ranges are slightly lower than what was calculated for 1991 alone in section 4.2.2. This is because the 1992 data included in the present calculation incorporate the low tussock emissions measured at T4-7 and also that the D sites were included in the wet tundra calculation. The former is thought to give a more realistic general tussock flux and the latter a better representativeness of wet tundra environments. It means, however, that the results are not directly comparable to Whalen and Reeburgh's estimates, as are those in section 4.2.2 and in the following section.

Assuming Alaska is representative for global tussock and wet tundra areas, the total emission should be in the order of 1.5-2.6 Tg/yr for tussock and 7.8-9.0 Tg/yr for wet tundras. In order to obtain global total tundra emission estimates by this approach similar data as in Table 4.4 would be needed for the other tundra subcategories. Such

data are not available and global flux estimates is therefore, for the time being, still based on more crude aproaches described in the following.

Table 4.4. Extrapolation of tussock and wet sedge tundra flux based on mean emission rates measured in 1991-92 on the North Slope of Alaska. The percentage cover of the different floristic units within tussock and wet meadow types of tundra was taken from Kummerow *et al.* (1983) and Walker *et al.* (1987). Alaska and global flux was based on a 100 day emission season.

Unit	Percent coverage	Mean flux ($mg/m^2/d$)	Unit flux ($mg/m^2/d$)	Landscape flux ($mg/m^2/d$)	Alaska flux (Tg/yr)	Global flux (Tg/yr)
Tussock tundra. Area: Alaska 0.126×10^{12} m^2. Global 0.992×10^{12} m^2.						
T	24-45	24	5.8-11			
C	3-7	150	4.5-11			
BH	30	1	0.3			
M	37-63	11	4.1-6.9			
Tussock				15-29	0.2-0.4	1.5-2.6
Wet meadow tundra. Area: Alaska 0.104×10^{12} m^2. Global 0.880×10^{12} m^2.						
D/C	80-90	109	87-98			
M	10-20	11	1.1-2.2			
Wet				88-100	0.9-1.0	7.7-8.8

4.4.2. Global extrapolation based on single studies

The approach taken in Table 4.4 is partly based on Whalen and Reeburgh's extrapolation scheme (Whalen and Reeburgh, 1988; Whalen and Reeburgh, 1992). This is a scheme designed for global extrapolation using the geographical data from digital databases discussed in Chapter 2. The scheme basically assumes Alaskan tussock tundra being representative of global dry/moist tundra which probably is open to

debate. However, the scheme was the first relatively detailed and widely used approach, and for comparative reasons it is also used to calculate global tundra flux in what follows. It also makes it possible to compare a particular year (1991) from Toolik Lake with the range of extrapolations found in the Fairbanks area.

Table 4.5. Global tundra CH_4 emission estimate. Methane flux is in g m^{-2} yr^{-1} and total emission is in Tg yr^{-1}. See text for sources.

Site type	Percent cover	CH_4 flux	CH_4 emission
	Dry/moist tundra (6.46 x 10^{12} m^2)		
Carex	3-7	8.83	1.7-4.0
Intertussock	30	0.067	0.1
Moss	37-63	1.24	3.0-5.0
Total dry/moist			12.1-22.9
	Wet meadow tundra (0.884 x 10^{12} m^2)		
Carex	80-90	8.83	6.2-7.0
Moss	10-20	1.24	0.1-0.2
Total wet meadow			6.3-7.2
Total tundra			**18.4-30.1**

Seasonal net flux for each unit was calculated by integrating the area under the mean flux curve using the trapezoidal rule. The possible effect on winter emission of difference in winter length between the North Slope and Fairbanks is assumed to be compensated by a thinner active layer in the north limiting the potential porewater reservoir in winter (Dise, 1992). Total tundra area was estimated as 7.34 x 10^{12} m^2 (Mathews, 1983) with 0.884 x 10^{12} m^2 wet meadow tundra (Mathews and Fung, 1987) and the remaining 6.46 x 10^{12} m^2 being moist tussock tundra (Chapter 2). Global tundra

CH_4 emission based on 1991 data from Toolik Lake is then calculated to be 18-30 Tg yr^{-1} (Table 4.5). The 1992 data from Toolik Lake, obtained partly in a very wet period, all show the highest fluxes measured anywhere in this study and as the calculation in section 4.2.1 indicates 1992 would seem to have produced a higher global estimate.

The estimates presented are a little lower than, but not conflicting with, the 35 Tg yr^{-1} that was suggested in a recent three-dimensional model synthesis of the global CH_4 cycle (Fung *et al.*, 1991) and they are within the overall range of 42±26 Tg yr^{-1} estimated by Whalen and Reeburgh (1992).

4.4.3. Literature extrapolation

Using the same crude subcategorisation of tundra as in the previous section, Table 4.6 shows results from a number of flux studies carried out in roughly ten different areas of the tundra and tundra-like environments north of 50°N. As can be seen from Figure 2.8 the studies have an Alaskan bias, due to the relative accessibility of tundra regions in Alaska.

The studies use different means of assessing representativeness of the two tundra types. For example, to estimate overall flux from the area, Bartlett *et al.* (1992) use a topographical transect while Whalen and Reeburgh (1992) and the present study use information on the areal coverage of different floristic units in each tundra type (as described above). All estimates are for the thaw season: Whalen and Reeburgh (1992) showed how winter emission in permafrost soils is insignificant on an annual scale. However, winter emission from non-permafrost northern wetland soils have been shown significant by Dise (1992) and the annual estimates from the southern part of the area considered here might therefore be slightly underestimated. On the contrary, most studies in Table 4.6 cover the high season, which might account for an overestimate of the seasonal integrated flux. Only the studies by Moore *et al.* (1990), Roulet *et al.*(1993), Whalen and Reeburgh (1992) and this study cover most of the thaw season. The longest study, by Whalen and Reeburgh (1992), covers four years' data of which the mean is given in Table 4.6.

Comparing the results in Table 4.6 directly is difficult . For example the wet tundra studies cover both ombrotrophic bog and minerotrophic fen environments with different soil pH. As will be discussed in Chapter 5, soil pH is known to be a limiting factor for methanogenesis, and similar fluxes should therefore not be expected in these environments. Likewise, as described above, the dry/moist tundra category covers a wide range of environments from the dry heath tundra to the moist tussock which often borders wet tundra. On a global scale, although representative for the sampling area, the upland tundra are probably too dry for an overall dry/moist average and the moist tussock tundra probably too wet. However, since the studies quoted in Table 4.6 cover wide ranges of environments, the data, taken together, are thought reasonably representative of these major tundra subdivisions.

The widest ranges of emissions are found in the longer time series studies of Whalen and Reeburgh (1992), Roulet *et al.* (1993), and the present study. In the time series there is a better chance of measuring occasional high emission. As described above, in Whalen and Reeburgh's four-year study, one whole year stood out because of an occasional very high flux from the *Carex* sites. This emphasises the point that time-series are important for providing more accurate seasonal flux estimates.

In Table 4.6 overall means of wet and dry/moist tundra were calculated in the same way as in Bartlett *et al.* (1992) but expanded by including the data presented above and this from Russian and Canadian studies. These extend the estimate to cover tundra environments north of 50°N including non-permafrost northern wetlands (Bartlett *et al.* only covers >60°N). The estimate of Bartlett *et al.* (1992) of 100±26 mg CH_4/m^2/day for wet tundra is very similar to the 96±20 mg CH_4/m^2/day calculated in Table 4.6. However, the dry/moist estimate, 11.5±3.5 mg CH_4/m^2/day, is higher (although not significantly) than the 5.8±1.4 mg CH_4/m^2/day calculated by Bartlett *et al.* (1992). This is not due to the added extra studies, but because a "mixed" zone of measurements in moist tussock tundra by Whalen and Reeburgh (1990b; 1992) included here, was excluded from the dry/moist tundra calculation by Bartlett *et al.* (1992). The weighting of measurements in dry and moist areas is obviously of great importance in the

calculation of global dry/moist tundra flux, but there are as yet no geographical information available that can quantify scales of moisture regimes and it is therefore not possible to estimate the exact contribution of "dry" versus "moist" tundra.

The assumption here is that each study has achieved reasonable representativeness of the tundra type it aimed to cover. Pooling all the studies in dry/moist and wet tundra areas should then represent a rough estimate of the global variety within each tundra type. However, it is acknowledged that ideally many more studies and more detailed knowledge about the areal extent of different moisture regimes are needed to make such extrapolations more reliable.

In Table 4.7 the mean fluxes are multiplied by the areal extent of each tundra type and the length of the thaw season to obtain an estimate for the annual global emission from these areas. The thaw season varies from about 160 days in the southern part of the area covered to about 80 days in the coastal marshes on the coast of the Arctic Ocean; and 120 days is thereforee chosen as a mean for the tundra as a whole. The emission from the wet tundra is much higher per areal unit than that from the dry/moist tundra. This means that, despite the large difference in areal coverage, the two types contribute about the same to the global figure of 19.5±5.1 Tg CH_4/yr.

This estimate could be somewhat of an underestimate due to a possible underrating of the episodic events mentioned earlier. On the other hand, as described above, the estimate might have been raised by the fact that most of the contributing data are peak season averages rather than means of an integration of all the thaw season. So there are great uncertainties in the figure. However, it does correspond with the conclusions drawn based on single field studies above, which show the tundra contribution to the atmospheric methane budget are somewhat lower than was generally thought throughout the 1980s (Fung *et al.*, 1991).

Table 4.6. Overall averages of thaw season northern wetland and tundra methane flux as measured in various field studies. Numbers in brackets in the first column refers to Fig. 2.7. Flux figures are in mg CH_4/m^2/day. Mean is ± standard error.

Habitat	**Flux**	**Range**	**Reference**
Wet tundra			
Wet communities (ombro/minero) (68°N) (11)	58.2	8.6-112	(Svensson and Roswall,1984)
Wet communities (minero) (68°N) (11)	360	80-950	(Svensson and Roswall,1984)
Wet coastal tundra (70°N) (5)	119	34-266	(Sebacher *et al.*, 1986)
Meadow tundra [a] (5)	40	9-78	(Sebacher *et al.*, 1986)
Sub-arctic fen (54°N) (10)	56.3	4.9-262	(Moore *et al.*, 1990)
Wet tundra [a] (4)	90	0-265	(Whalen and Reeburgh, 1990b)
Wet meadow tundra (64°N) (7)	110	0-2216	(Whalen and Reeburgh, 1992)
Wet meadow tundra (68°N) (6)	97	2-1500	This study
Depressions (68°N) (13)	138 [b]	25-251	(Panikov and Zelenev, 1992)
Ombro. bog (56°N) (12)	166	-	(Panikov and Zelenev, 1992)
Coastal marsh (70°N) (2)	52	0-150.7	(Samarkin *et al.*, 1992)
Low polygonal ground [a] (3)	46.1	-	(Morrissey and Livingston, 1992)
Meadows [a] (3)	64.4	-	(Morrissey and Livingston, 1992)
Wet meadow (60°N) (1)	29	-	(Fan *et al.*, 1992)
Wet meadow (60°N) (1)	144	16-426	(Bartlett *et al.*, 1992)
Wetland (51°N) (8)	16	-2-1626	(Roulet *et al.*, 1993)
Wetland (58°N) (9)	44	-3-2255	(Roulet *et al.*, 1993)
Mean	**95.9±19.6**		
Dry/moist tundra			
Ombro. communities (68°N) (11)	11.6	0.3-29	(Svensson and Roswall, 1984)
Moist tundra [a] (5)	4.9	0.3-12	(Sebacher *et al.*, 1986)
Alpine tundra [a] (4)	0.6	-0.2-6.3	(Whalen and Reeburgh, 1990b)
Moist tundra [a] (4)	31	0-159	(Whalen and Reeburgh, 1990b)
Peat hills and base (68°N) (13)	6 [b]	0-12	(Panikov and Zelenev, 1992)
High polygonal ground [a] (3)	4.9	-	(Morrissey and Livingston, 1992)
Tussocks [a] (3)	3.4	-	(Morrissey and Livingston, 1992)
Inter tussocks [a] (3)	2.9	-	(Morrissey and Livingston, 1992)
Moist tundra (64°N) (7)	35	0-2216	(Whalen and Reeburgh, 1992)
Moist tundra (68°N) (6)	25	0-1500	This study
Upland tundra (60°N) (1)	2.3	-2.1-18	(Bartlett *et al.*, 1992)
Upland tundra (60°N) (1)	11	-	(Fan *et al.*, 1992)
Mean	**11.5±3.5**		

[a] indicates north-south transect studies.
[b] only range given. Mean taken as average of range.

Table 4.7. Global tundra (>50°N) methane emission estimate based on literature values of average flux in Table 4.6. Flux is in mg $CH_4/m^2/day$.

Tundra type	Area x 10^{12} m^2	Average flux ±st. error	Thaw season days	Global emission Tg CH_4/yr
Wet	0.884	96±20	120	10.2±2.1
Dry/moist	6.460	12±4	120	9.3±3.0
Total				**19.5±5.1**

4.4.4. Total northern wetland and tundra flux

The global estimates suggested above of tundra and non-forested wetland emission from latitudes >50°N (18-30 and 19.6±5.1 Tg CH_4/yr based on field survey and literature average respectively) do not include emission from forested bogs, lakes and possible significant winter emission from southern non-permafrost areas. The forested bogs cover 1.77 x 10^{12} m^2 (Mathews and Fung, 1987). An extensive study including flux from forested bogs north of 50°N (discussed above, Roulet *et al.*, 1993) suggests these environments having a maximum flux of 5 mg $CH_4/m^2/day$, which would yield a global figure of 1 Tg CH_4/yr, given a 120 day active period. The large regional differences in boreal flux mentioned earlier have given rise to higher estimates for forested bog flux (Bartlett and Harriss, 1993). However, the higher fluxes were nearly all measured in areas south of 50°N and are therefore excluded from the present extrapolation.

Small and shallow lakes are assumed to have significant emissions. Roulet *et al.* (1993) found a mean of 125.5 mg $CH_4/m^2/day$ in shallow lakes of the Hudson Bay Lowland and Bartlett *et al.* (1992) found an average of 77 mg $CH_4/m^2/day$ in small lakes and ponds of sub-arctic Alaskan tundra. Kling *et al.* (1992) found lower but still significant fluxes of 1-16 mg $CH_4/m^2/day$ from large lakes on the North Slope of Alaska. Bartlett *et al.* (1992) also found relatively small fluxes (3.8 mg $CH_4/m^2/day$) in larger lakes. However, geographical data on the coverage of lakes in general, and the

ratio of small to large lakes in particular, are presently lacking. It is therefore not possible to estimate global >50°N lake emission accurately. Added together, lake emission, unaccounted-for winter emission (Dise, 1992), and 1 Tg CH_4/yr for forested bog emission is thought unlikely to exceed 5 Tg CH_4/yr. The overall tundra and northern wetland (>50°N) emissions is therefore calculated at 19-35 Tg CH_4/yr.

Fung *et al.* (1991) used a three-dimensional model to simulate the sizes of global sources and sinks of atmospheric methane. Their preferred scenario arrived at 35 Tg CH_4/yr for total northern wetland and tundra flux. Similarly, Bartlett and Harriss (1993) and Harriss *et al.* (1993), reviewing literature data on methane emission from high latitude wetlands, estimate that between 34 and 38 Tg CH_4 are emitted north of 45°N. These estimates are at the maximum of the range calculated here partly due the differences in forested bog flux mentioned earlier.

In the same study isotopical data was used to constrain the budget scenarios. In the preferred scenario a $\delta^{13}C$ ratio of -61‰ was chosen for tundra emission of CH_4. This is also in agreement with the results of isotopical analysis of CH_4 emitted from my tundra sites, which shows a mean $\delta^{13}C$ of -63.8±3.9 ‰ (see section 3.4 and Table 5.3).

Future improvements of global tundra emission estimates depend on more field surveys of the large northern wetland areas in Siberia, in particular, and undoubtedly also on a more extensive use of remote sensing for extrapolation purposes. Landsat Thematic Mapper images have been used for providing probably the most accurate regional estimates of methane flux from wetlands (Bartlett *et al.*, 1989; Roulet *et al.*, 1993). However, the use of such images for global extrapolations would be prohibitively expensive. A "back-of-the-envelope" calculation suggests (just for the images) a price of approximately £0.25/km^2 or £1.8 million for such a study of global tundra areas. Regional studies using remote sensing will provide useful and more precise information to compare with the less accurate "multiply-flux-by-x km^2"extrapolations.

4.5. Summary

Six floristic units representing of arctic tundra were shown to have significantly different seasonal mean fluxes ranging from 112±71.8 mg CH_4/m^2/day to 0.6±0.4 mg CH_4/m^2/day. The rank order from high to low flux was: *Carex*, depressions, tussocks, elevations, mosses and inter-tussock depressions. In general, the wetter the soil and the more vascular plants present, the higher the flux. Episodic events of very high flux were repeatedly measured and may account for up to 25% of the annual tundra flux to the atmosphere.

Using information on the areal coverage and the 1991 seasonal mean flux from the floristic units, an overall flux of 97±6 mg CH_4/m^2/day for wet meadow tundra and 25±8 mg CH_4/m^2/day for moist tussock tundra was calculated. Similarly, but when using mean of all measurements in 1991 and 1992, Alaskan tussock tundra emission were calculated at 15-29 mg/m^2/day and wet tundra emission at 88-100 mg/m^2/day. This yields total figures for Alaska of tussock tundra flux 0.2-0.4 Tg/yr, and wet tundra 0.9-1.1 Tg/yr. Assuming Alaskan tundra is representative for global tundra this results in global fluxes of 1.5-2.6 Tg/yr for tussock and 7.8-9.0 Tg/yr for wet tundra. Since most data was obtained in Alaskan tussock and wet tundra environments these are probably the most confident extrapolations that can be produced based on the presented data.

For comparative reasons a widely used scheme for extrapolating to global tundra methane fluxes was also employed. An overall tundra flux from the North Slope of Alaska weighted by the relative global coverage of wet and moist tundra as assumed by this scheme yields 26-41 mg CH_4/m^2/day. Based on these data and this extrapolation scheme, a global tundra methane emission is calculated at 18-30 Tg CH_4/yr. An attempt to pool all available literature studies in order to better represent regional differences in tundra flux yields a global figure of 19.5±5.1 Tg CH_4/yr. Combining these two estimates with fluxes from remaining parts of all northern wetlands and tundra north of 50°N derives a total estimated flux of 19-35 Tg CH_4/yr.

Chapter 5

Controls on methane emission from tundra soils

5.1. Introduction

The methane produced at depth in tundra soils is subject to varying degrees of microbial oxidation in the upper soil layers, and net emission is a function of the balance between these two processes (Figure 3.2). Net emission is dependent upon the effect of temperature, moisture and other factors limiting these two biological processes, as well as on physical factors limiting gas transport in the soil.

In this chapter I analyse how the net methane emission is controlled by environmental factors. Following from my presentation of the scale of emission found in the field (Chapter 4) I first describe and discuss field results (section 5.2 and 5.3) supported by some observations in the laboratory (section 5.4). These two sections focus on the effect of soil temperature, thaw depth and moisture. Subsequently I give an overview of the known controlling factors and discuss which factors are of major importance considering changes in climate.

5.2. Temporal variations in flux at Toolik Lake

Methane flux, soil temperature and water table height as measured at the different floristic units through the 1991 and 1992 seasons are illustrated in Figure 4.4. Table 5.1 summarises general soil environment data obtained during the 1991 thaw season. Thaw depth developed at all sites in 1991 from 10-30 cm below soil surface when sampling started on June 15 to the maximum depth (20 - 70 cm) before the end of June. This depth was maintained at most sites until sampling stopped on September 1. Only the moss sites experienced surface freeze-up before that, on August 28.

At T, BH and M sites the maximum thaw depth dropped below the organic horizon and into the mineral soil. The mineral soil is thin at these sites since the most recent glacial advance from Brooks Range, "Itkillik II" (about 11,000 yr BP), covered

the area (Hamilton and Porter, 1975; Hamilton, 1986). This means bedrock is at a shallow depth and in summer at sites such as T4-7 bedrock is reached before permafrost.

Table 5.1. Soil environment data from different floristic units at sites near Toolik Lake as observed between June 15 and Aug 30 1991. Temperatures in °C are measured as an average (± range) of the top 13 cm of the soil. Water table position is measured in cm (mean ± range) relative to soil surface (negative value being below soil surface). Thaw depth and depth of the organic layer are measured in cm below soil surface. Soil pH was measured in slurries of top 10 cm soil.

	BH	M	T	C	D	E
Temperature						
mean (°C)	2.3±2.4	3.4±3.1	7.6±6.1	8.0±7.2	7.3±4.1	9.7±6.3
Water table						
mean (cm)	14.1±7	-9.2±12	-13.5±9	-4.1±13	5.7±5	-11.8±6
Thaw depth						
mean (cm)	27.5	16.9	63.5	47.6	44.4	62.6
min.	12	0	19	31	23	37
max.	30	20	70	51	50	70
Organic layer depth (cm)	<10	10	>30	>40	>30	>30
pH	5.4	5.4	4.5	4.9	4.5	4.5

The flux and soil environment data presented here seem generally to support Whalen and Reeburgh's (1992) recent conclusion that single-parameter relationships used to predict methane flux are site-specific. However, correlations with soil temperature and water table at the *Carex* site seems to indicate a possible rank order of controls where, once the water table drops below the soil surface and allows increasing rates of microbial oxidation to occur, a simple temperature/flux relationship is overruled (Figure 4.4.a). At the C sites the water table gradually declined throughout the 1991 season, and around July 1 it dropped below the soil surface (Figure 4.4.a). If

the episodic event at C2 (see Chapter 4) is excluded and the two periods are considered separately, a linear correlation with temperature (r^2=0.79, n=6) is seen in the waterlogged period. When the water table dropped below the soil surface the flux decreased, correlating logarithmically with the falling water table (r^2=0.84, n=14). A very similar correlation between drying of *Carex* sites and decreasing emission was reported by Bartlett *et al.* (1992).

These results suggest that on a short-term basis temperature might independently control emission at the waterlogged sites, while moisture controls emission from sites where the water table is below the soil surface. This could explain the correlations found, and interaction between temperature and moisture might account for lack of relationships when temperature is related to all-season flux data from dry/moist sites. However, such simple explanations are challenged by the complexity of controlling factors found by Whalen and Reeburgh (1992) and by results of Bartlett *et al.* (1992), who found a good correlation between soil temperature and flux at dry upland tundra sites. Taking into account the complexity of controlling factors, particularly when the water table is below the soil surface, simple correlations between all-season flux and single environmental parameters should not be expected. However, floristic units in a constantly wet environment could be expected to show better correlation between soil temperature and flux relative to dryer sites assuming that high moisture content limits oxidation. This is not entirely the case although in the early season (mid to late June) the emissions at all sites generally increased with the progressive warming of the soil. A period of very cold weather in early July 1991 caused a dramatic fall in soil temperatures and was followed by reduced flux at most sites, in particular the wet C, D and E sites (Figure 4.4.a-c). In spite of this, only C sites showed a significant positive association between soil temperature and flux when all-season data are compared using Spearman's rank correlation test (Table 5.2).

The reason for absence of statistically significant relationships is that the same temperature generally corresponds with higher flux at the end of the thaw-season compared to measurements at the start of the season (Figure 4.4.b, c, e, f), a pattern also

found by Svensson and Roswall (1984). One way of explaining this is by an increasing "background" emission which is not influenced by the immediate soil environment but by the progressing season. Simple explanations for this background emission could be increased microbial population size, an increasing supersaturated content of methane in the soil water, or a combination of these factors. However, whatever the course, the consequence is an increased emission throughout the summer.

Table 5.2. Spearman's rank correlation (r_s) coefficient for correlations between CH_4 emission and environmental variables. Only values of r_s significant at $p<0.05$ are shown.

	C	D	E	BH	M	T
n	20	20	21	20	19	20
Soil temp.	0.52					-0.46
Water table					0.57	0.54

It is hypothesized that this background emission may be quantified as an increasing function of degree days, and if this factor is subtracted from the measured flux a better measure of production over time is provided. An arbitrary "pool" is calculated by subtracting start-season from end-season emissions at the same soil temperatures and the background emission is calculated in the following way:

$$p = f_{end} - f_{start}$$

$$b_i = \frac{p \times dd_i}{dd_n}$$

$$cf_i = mf_i - b_i$$

where p is the "pool" and f_{start} and f_{end} are flux at equivalent temperatures at the start and end-season, respectively. Background emission (b_i) is equal to the fraction of degree days on the day in question (dd_i) to all season degree days (dd_n) multiplied by p. The corrected flux (cf_i) is then calculated by subtracting the background emission from the actual measured flux (mf_i).

Below follows a sample calculation of corrected flux on julian day 195 (July 14) at the D sites: f_{start} =17.8 mg/m^2/day (June 15); f_{end} = 60.3 mg/m^2/day (Aug. 30); dd_{195} = 293; dd_n (Aug. 30)=619; mf_{195} = 64.2 mg/m^2/day.

$$p = 60.3 - 17.8 = 42.5 \text{ mg m}^{-2}\text{day}^{-1}$$

$$b_{195} = \frac{42.5 \times 293}{619} = 20.1 \text{ mg m}^{-2}\text{day}^{-1}$$

$$cf_{195} = 64.2 - 20.1 = 44.1 \text{ mg m}^{-2}\text{day}^{-1}$$

Table 5.3 show how linear relationships for all data comparisons between flux and soil temperature change to become statistically significant at the wet D and E sites when the data are corrected as described. Looking at the raw data from the D and E sites (Table 5.3), the first period (June 15-July 6) with increasing soil temperature shows substantially higher regression slopes than the last period (Aug. 1-Aug. 30) with decreasing temperature (slopes: 7.2 and 2.4 over 4.6 and 1.65, respectively). This is in accordance with the theory which implies that the flux will be more strongly affected by increasing than by decreasing temperatures due to the increasing pool. Also this difference between slopes should diminish when the data are corrected. This is in fact what tends to happen at the D sites (from 7.2 and 4.6 to 6.7 and 6.0; Table 5.3). The corrected data for the last period at the E sites could not be compared due to a shortage of data. A limited number of observations at the end of the season could not be used when calculating the correction at E sites because the calculation required equivalent start- and end-season temperatures. This could be obtained at the E sites only by excluding the last two data points in the season.

Whalen and Reeburgh (1992), measuring p_{CH4} distributions in soil water at their sites in the summers of 1988 and 1989, found no clear relationships between p_{CH4} and CH_4 flux. Their sites were generally dryer than the two used for the correction above, but 1988 was relatively wet and best suited for comparison with the sites in this study. That year Whalen and Reeburgh observed an increase in soil p_{CH4} over the season

followed by dramatic late season increases in CH_4 flux, which could support the theory described above.

Table 5.3. Regression analysis for the association between methane flux and mean soil temperature at Depression and Elevation sites near Kuparuk River in 1991. Comparisons of both raw and corrected (Cor.) data as described in the text are shown. Temperature was increasing consistently in the period between June 15 and July 6. It was decreasing between Aug. 1 and Aug. 30. Probability level, p, is from an F test with * = $p<0.05$, ** = $p<0.01$ and *** = $p<0.001$. NS: Not significant.

	Slope		r^2		p	
	Raw	Cor.	Raw	Cor.	Raw	Cor.
D sites						
June 15-July 6	7.2	6.7	0.76	0.79	**	**
Aug. 1-Aug. 30	4.6	6.0	0.83	0.88	**	**
All season	3.5	5.3	0.18	0.62	NS	***
E sites						
June 15-July 6	2.4	2.5	0.55	0.80	NS	**
Aug. 1-Aug. 30	1.7	-[a]	0.49	-[a]	NS	-[a]
All season	0.6	1.2	0.06	0.41	NS	**

[a] Too few observations available (see text).

To test if the described pattern observed in the field could be reproduced in a controlled environment, an experiment were carried out which will be dealt with in section 5.4.

Both M and T sites show significant non-parametric correlations with water table (Table 5.2) which corresponds with the hypothesis that at these drier sites the water table becomes the primary control on emission by determining the rate of methane oxidation.

In August 1992 a transect of T sites was established at 50 meter intervals up-slope from the Kuparuk River. The transect showed a gradual decrease in CH_4 emission when moving from the moist T1, with the highest water table to T7 with the water table below bedrock (Figure 5.1). T7 showed a small rate of CH_4 consumption when it was

first sampled on 7 August. Following heavy rainfall and a water-table rise of 15 cm, the consumption stopped but resumed when the water table had fallen again (Figure 5.2). The observations at the transect give further evidence for a strong water table control on CH_4 flux in tussock environments and also show the potential for dryer tundra areas to act as an atmospheric CH_4 sink.

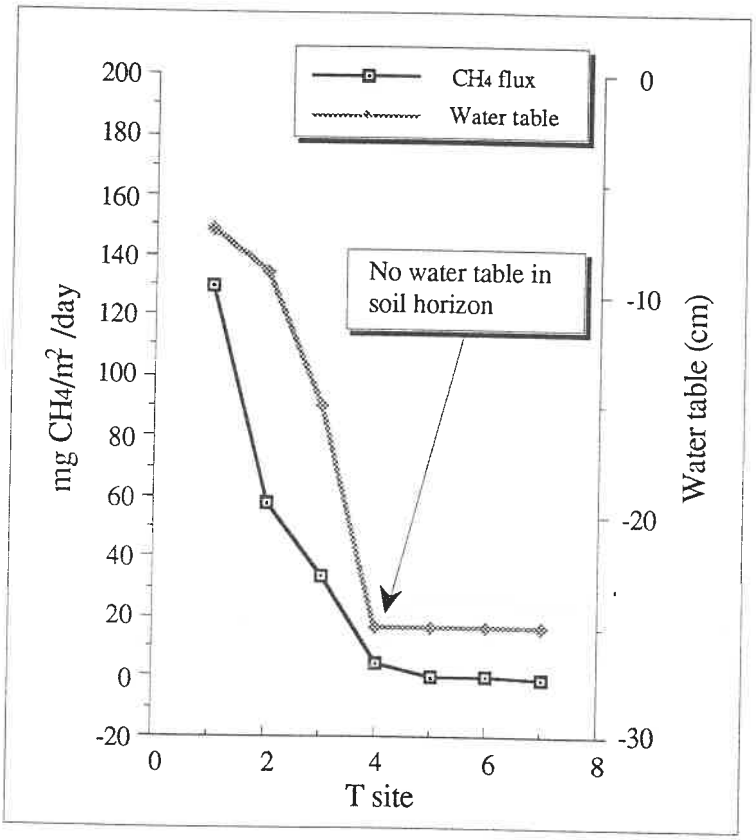


Figure 5.1. Methane flux and water table position at seven T sites on a transect with approximately 50 meter intervals up slope from Kuparuk River on 7 August 1992.

The 1991 data from the T sites also show a smaller inverse correlation with temperature (Table 5.2). This can be biologically reasonable only if methane-oxidising bacteria are highly sensitive to the decreasing temperature while methane producing bacteria are not. The methane-oxidising bacteria are more exposed to temperature changes simply because the aerobic zone is closer to the soil surface, but laboratory studies on temperature dependency of the two groups of bacteria (from equivalent soil environments) are needed to determine whether their responses differ in a way that can

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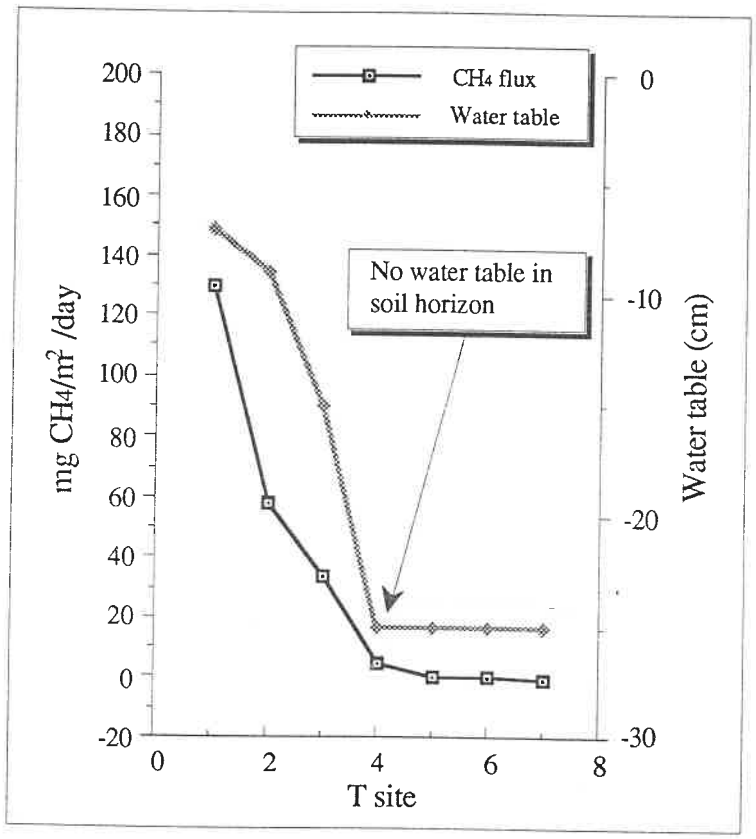

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account for the possible relationship seen here. BH showed no significant relationship with environmental parameters.

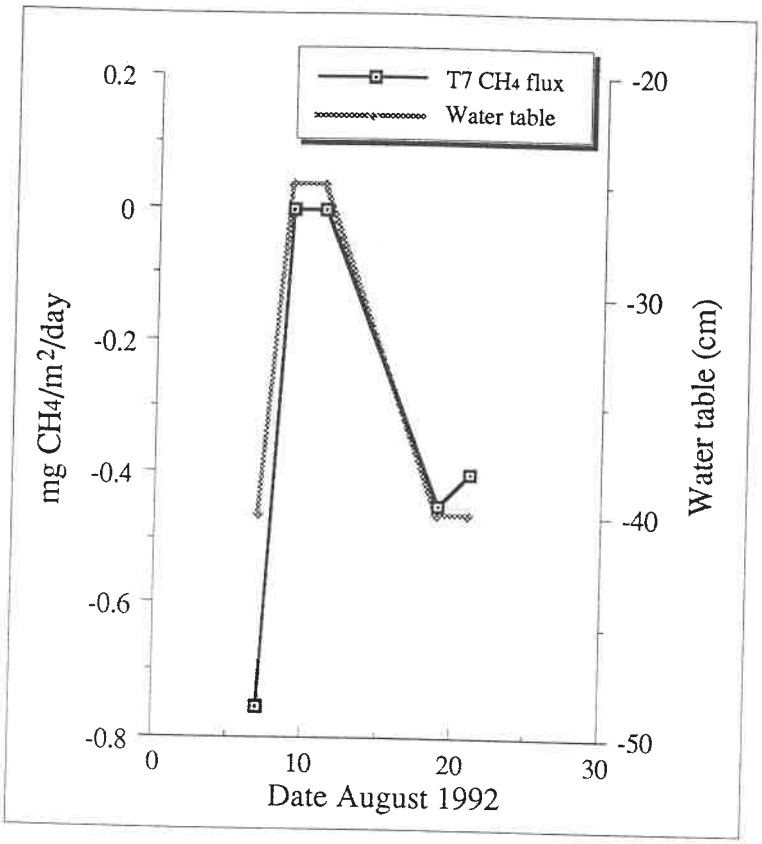


Figure 5.2. Methane uptake by site T7 near Kuparuk River as observed during August 1992. The rise in water table around 10 August was from being non-existent in the soil horizon (here illustrated as -40cm) to a position just above bedrock.

In general, methane flux at the moist and dry upland tundra sites, T and M, seems to be controlled primarily by the level of the water table, while flux at the wetter meadow sites, D, E and C (when inundated) seems to be controlled mainly by temperature. It is assumed that at the wet sites the observed relationship is reflecting temperature dependency of the methanogenic bacteria, which is consistent with results from earlier studies (Svensson and Roswall, 1984; Crill *et al.*, 1988; Moore *et al.*, 1990; Bartlett *et al.*, 1992). When the water table drops below the soil surface, the microbial methane oxidative activity in the aerobic soil layer becomes the key biological factor controlling net flux to the atmosphere. The limitation of the oxidation can be quantified

as the position of the water table, although it has not been shown that it is the actual aerobic space which is controlling the methane oxidation. However, moisture has been identified as limiting soil microbial oxidation of CH_4 in many studies (Steudler *et al.*, 1989; Yavitt *et al.*, 1990; Bartlett *et al.*, 1992) and a significant effect of microbial CH_4 oxidisers on net CH_4 flux from sub-arctic tundra soils has also been shown (Whalen and Reeburgh, 1990a) (see Chapter 3).

5.3. Isotopical signature of tundra methane

Isotopical analyses of the $^{13}C/^{12}C$ and $^{14}C/^{12}C$ ratios were carried out on CO_2 and CH_4 and in addition also the D/H ratio in CH_4 on samples from sites which had sufficient high emission to allow accumulation of enough gas to carry out the analyses (that is, C, D and T sites). This work was carried out in collaboration with Dr. Ingeborg Levin and Marcus Thom at the Institut für Umweltphysik, University of Heidelberg in Germany (see Preface). Samples were taken in August 1992 and shipped to Heidelberg for analyses. Carbon-14 analyses were carried out by ETH Zürich in Switzerland. Details of techniques for sampling and analysis can be found in Appendix 1.

The main objective of the isotope work was firstly to get information on the general isotopic signature of the emitted gases. Secondly, the idea was to look more closely at differences in isotopic composition between stirred (bubble, unoxidised) methane and methane that was assumed to have been exposed to partial oxidation ("diffusive" methane, see section 3.4).

Weighted $\delta^{13}C$ means of the decomposing plant material, as derived from $\delta^{13}C$ measurements in the emitted CO_2 assuming no fractionation during aerobic decomposition of plant material and no CO_2 contribution from CH_4 oxidation, are -24.8‰ for *Carex* and -25.3‰ for depression sites. These values are in the expected range for terrestrial C_3 plants (-32 to -23‰) (Ehleringer and Rundel, 1988). Tundra does not contain C_4 plants which are isotopically enriched in ^{13}C. All terrestrial tundra plant species should therefore have $\delta^{13}C$ values near the C_3 mode.

Few published data are available on the isotopic compostion of methane emitted from tundra. The mean "diffusive" $\delta^{13}CH_4$ of -63.9±3.9‰ in Table 5.4 corresponds well with the -65.8±2.2‰ found in a major study characterising isotopic composition of methane emitted from tundra environments in the Yukon-Kuskokwim Delta (Martens *et al.*, 1992).

No significant differences between means of "stirred" and "diffusive" samples were found. The isotopic fractionation carried out by wetland plants, as shown by Chanton *et al.* (1992a), is capable of outweighing the fractionation caused by oxidation, and thus could cause the inconclusive result.

	$\delta^{13}CH_4$ (‰)	δD (‰)	$\delta^{13}CO_2$ (‰)	$\delta^{14}CH_4$ (pMC)	$\delta^{14}CO_2$ (pMC)
Carex	-56.19 (2)	-264 (2)	-20.9 (2)	104.7 (2)	112.6 (2)
stirred	-58.2 (1)				
Depres.	-68.27 (1)	-317 (1)	-19.66 (1)	109.2 (1)	114.8 (1)
stirred	-64.55 (1)	-340 (1)		110.7 (1)	
Tussock	-74.38 (1)		-24.83 (1)	113.7 (1)	
Mean	-63.9±3.9	-290	-21.6±1.1		
stirred	-61.4				

Table 5.4. Summary of isotopic data from analysis of methane and carbon dioxide emitted from tundra sites on the North Slope of Alaska. Site descriptions are in section 4.2. See Appendix 1 for details of methods for sampling and analysis. Figures in brackets are number of samples. See section 3.4 for explanation of units. Note that contamination by stray water have caused some uncertainty about the deuterium values.

The data when plotted in a $^{13}C/D$ scheme such as Figure 3.4 cover a wider area than quoted by Wahlen (1993). The plot produced by Wahlen suggests that by stable isotope measurements it should be possible to distinguish tundra methane from, for example, other wetlands and ruminants. Our data seem to indicate that, although there

is uncertainty about the deuterium values (see Appendix 1), the method might not be quite as unambiguous as proposed by Wahlen (1993).

The isotope data (Table 5.4) indicate that, in fact, it is oxidation that increases as a function of lowering water table. The experiments were carried out when the water table was below the soil surface at the *Carex* sites and the Depression sites were, as always, inundated. At the Depression bubble methane and diffusive flux show an anticorrelation between ^{13}C and D ($\Delta D/\Delta^{13}C = -5$). This could be due to different contributions from acetate fermentation relative to CO_2 reduction between the two locations of sampling (a couple of meters apart). However, it cannot be due to any degree of oxidation of the diffusive methane since this process would lead to positive correlation of the two isotopes (see section 3.4). It is therefore assumed that methane emitted at the depressions have not been partly oxidised. Correspondingly, diffusive fluxes were enriched at the *Carex* site relative to Depression both in terms of ^{13}C and D with a $\Delta D/\Delta^{13}C$ ratio of approximately 7. This indicates partial oxidation of methane in the soil at the *Carex* site since fractionation factors for oxidation are in the range 8-14 (Coleman *et al.*, 1981). This supports the assumption that aerobic methane consumption is limited by the position of the water table.

It is unfortunately not possible, using the ^{14}C data, to determine the exact age of the decomposing plant material. The reason is the nuclear bomb contamination of atmospheric ^{14}C mentioned in section 3.4. Decaying organic material is assumed to be composed by three "reservoirs", a young (Y), middle (M) and old (O) age. The turnover times of these reservoirs are assumed to be 1, 30 and 1000 years respectively. The decay can then be simulated by assuming Y emitting the mean atmospheric ^{14}C value of the previous two years, M the mean value of the previous 60 years and O the mean value of the past 2000 years. There is obviously many ways in which the mixing of Y+M+O can be composed. Taking the 105 pMC of methane emitted from the *Carex* site and making a simple calculation based on atmospheric ^{14}C data for the past 200 years yields a mean value of the past 180 years (Marcus Thom, pers. comm. 1993). However, to say the source is 90 years old is not correct because of the unknown

contribution from the Y reservoir, which shifts the value towards the present atmospheric $^{14}CO_2$ value of 114 pMC (Levin *et al.*, 1992) but it does seem the source is more than just a few decades old.

5.4. Laboratory experiments

A soil core experiment was carried out to test the temperature dependency of northern wetland emission in a controlled environment. Soil cores of 45 cm depth were collected in Bøllemosen (55°50'N, 12°36'E) an acid dystrophic bog 15 kilometres north of Copenhagen, Denmark. Four cores were taken at floristically similar sites in the central non-forested open part of the bog and four cores were taken in the forested bog margin.

The cores were brought to a laboratory at the Institute of Population Biology, University of Copenhagen, and incubated initially at 5°C (temperature similar to the bog soil when cores were taken) in a temperature controlled environment. The cores were kept inundated through a simulated thaw season, bringing them stepwise to 10, 15 and 20°C and similarly back again to 5°C. Details of the methodology are found in Appendix 1.

Figure 5.3 shows temperature and flux with time for open bog (a) and forested margin (b). The mean scale of emission was discussed in section 4.3. There is a clear but somewhat delayed response to the increasing temperatures in the open bog cores. Following the degree day hypothesis outlined in section 5.2, a higher flux at equivalent temperatures would be expected when temperatures came back down after the peak 20°C. This is also what happened in the bog cores (Figure 5.3.a). However, there is a problem in validating the theory on the basis of this experiment since the timing was different when the temperatures were increasing from when they decreased (due to an unfortunate limited access to the gas chromatograph where the samples were analysed). However, the concentration of dissolved methane in the pore water did increase during the same period (Figure 5.4.a), which seems to indicate that part of the observed difference in flux before and after the peak temperature could be due to an increase in physical release from the pore water. The bog cores showed no significant pattern in

pore water concentration with depth (mean ± standard error at 12.5 cm depth: 10.8±2.7 mM; 25 cm: 12.9±3.4mM and 37.5 cm: 13.2±3.0 mM).

As mentioned in section 4.3, the forested bog margin had generally very low flux. At 20°C it suddenly emitted about 100 times the normal level but returned quickly to around 1-3 mg/m^2/day (Figure 5.3.b). Under natural conditions the soil temperatures will not rise above 15°C and the high emission is therefore probably unrealistic *in situ*.

The difference in pore water concentration between bog and forest cores is probably not significant since the relative water contents of the two soil types are very different. The water holding capacity of inundated forested bog margin and open bog cores, calculated as g H_2O/100 g dried soil, were 943 and 2537 respectively, or almost three times as high in the open bog. A rough indication is therefore that any dissolved methane concentration in the forested margin corresponds to one third of it in the open bog.

In general (despite the problems with timing the experiment) the open bog core flux and porewater concentration seem to support the theory that a steady increase in the latter with the progressing season could support part of the higher flux observed at the end of the season. The forested bog soils seems, despite the low flux, to support a more intense cycling of methane between microbial producers and consumers.

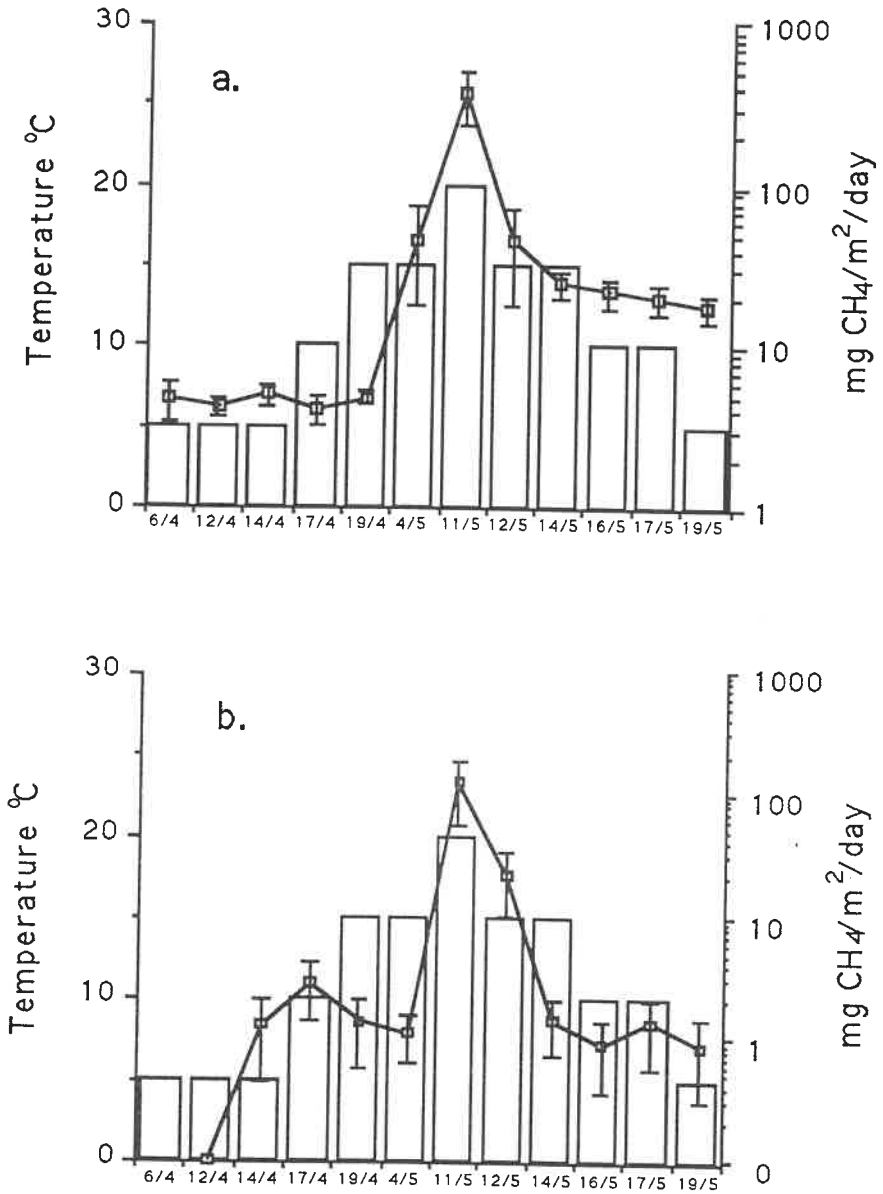


Figure 5.3. Mean flux ± st. error (n=4) and soil temperature (bars) in cores taken from a: an open bog and b: a forested bog margin, in Bøllemosen, Denmark.

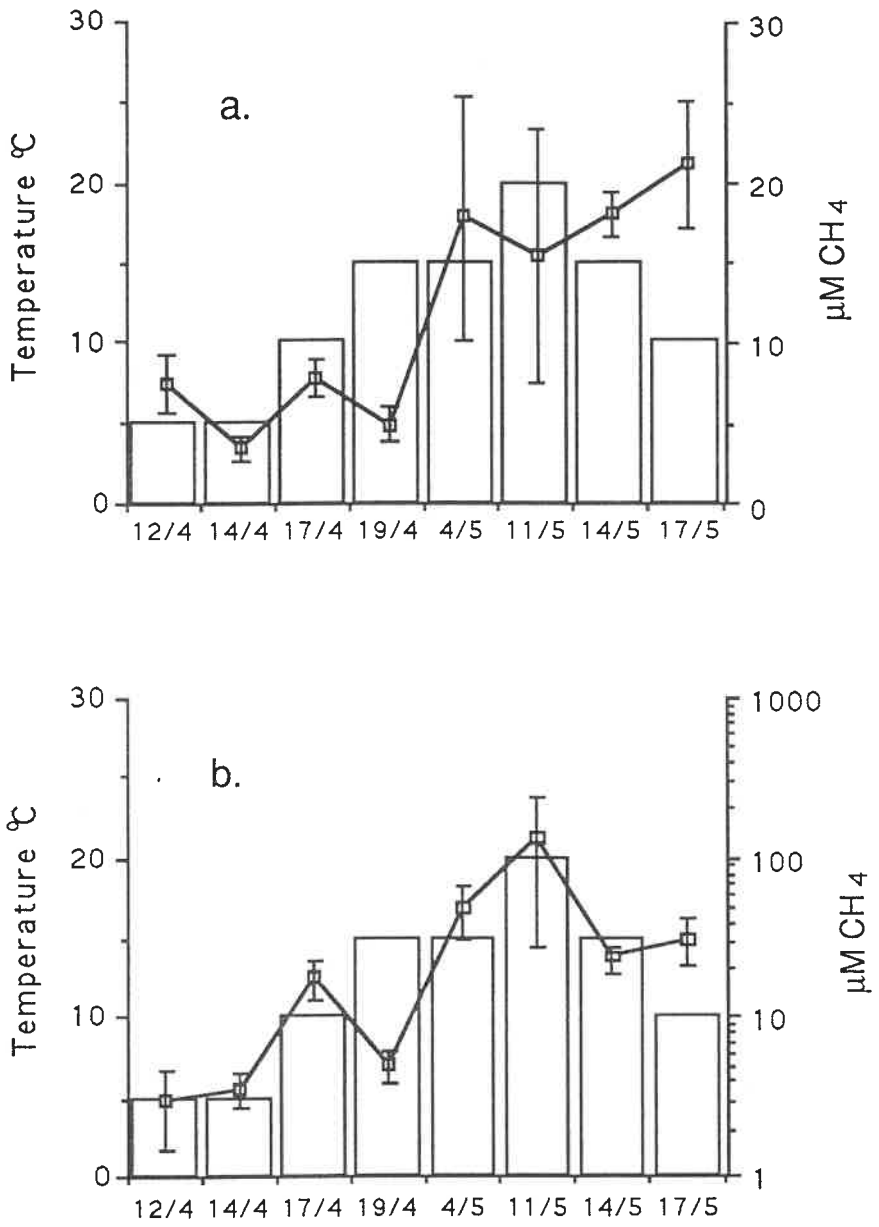


Figure 5.4. Mean dissolved methane concentration (mean of three depths: 12.5 cm, 25 cm and 37.5 cm in four cores) and soil temperatures (bars) in cores from a: open bog and b: forested bog margin.

5.5. Discussion

In this section I overview most of the factors suggested that are to control tundra methane flux. It will partly form a summary of the data presented above. Emphasis will be on evaluating the relevance of the different factors in terms of modelling attempts, like the one to be described in Chapter 6.

5.5.1. Temperature

Probably the most obvious sign of a temporal temperature control on methane emissions is the strong seasonality shown clearly in multi-year studies (Figure 5.5). Increasing emission follows the spring thaw of the soil very closely. The reason for this correlation is probably a gradual release of methane trapped in the frozen soil over winter. However, the microbial communities are also increasing in activity and population size with warming of the soil and deepening of the active layer. The nature of the correlation between net emission of methane and soil temperature has been observed to be either linear (Svensson and Roswall, (1984); this study), logarithmic (Crill *et al.*, 1988; Moore and Knowles, 1990; Bartlett *et al.*, 1992), or exponential (Moore *et al.*, 1990). As shown above, the relationship is a function of the effect of temperature on both CH_4 production and consumption and is, hence, not straightforward. Any dependency of the CH_4 flux on simple environmental factors should therefore not be expected in areas where the water table is below the soil surface and both processes are operating. However, in constantly inundated soils where oxidation is thought to be at a minimum, the net emission might become a more simple function of soil temperature and thaw depth. Whalen and Reeburgh (1992) developed a value called centimeter-degrees (a product of thaw depth and mean soil temperature from surface to permafrost) that showed the best non-parametric correlation with net methane emission. Svensson and Rosswall (1984) and the present study both found stronger relationships between net emission and soil temperature in constantly wet as opposed to dryer sites. However, as described above, it was also found that the correlation was lower when calculated over a whole season than part of a season due to

increasing rate of emission as the season progressed. Some form of correction of the flux/temperature relationship as described in section 5.2 is therefore necessary to incorporate in the calculations when establishing predictive models of methane flux.

From a modelling perspective the temperature effect on net methane emission from tundra environments is of crucial importance. However, temperature/flux relationships are not straightforward, and effects on production and consumption processes need integration when incorporated into models. Also the potential effect of a changing temperature regime on the active layer depth is of great importance (see below).

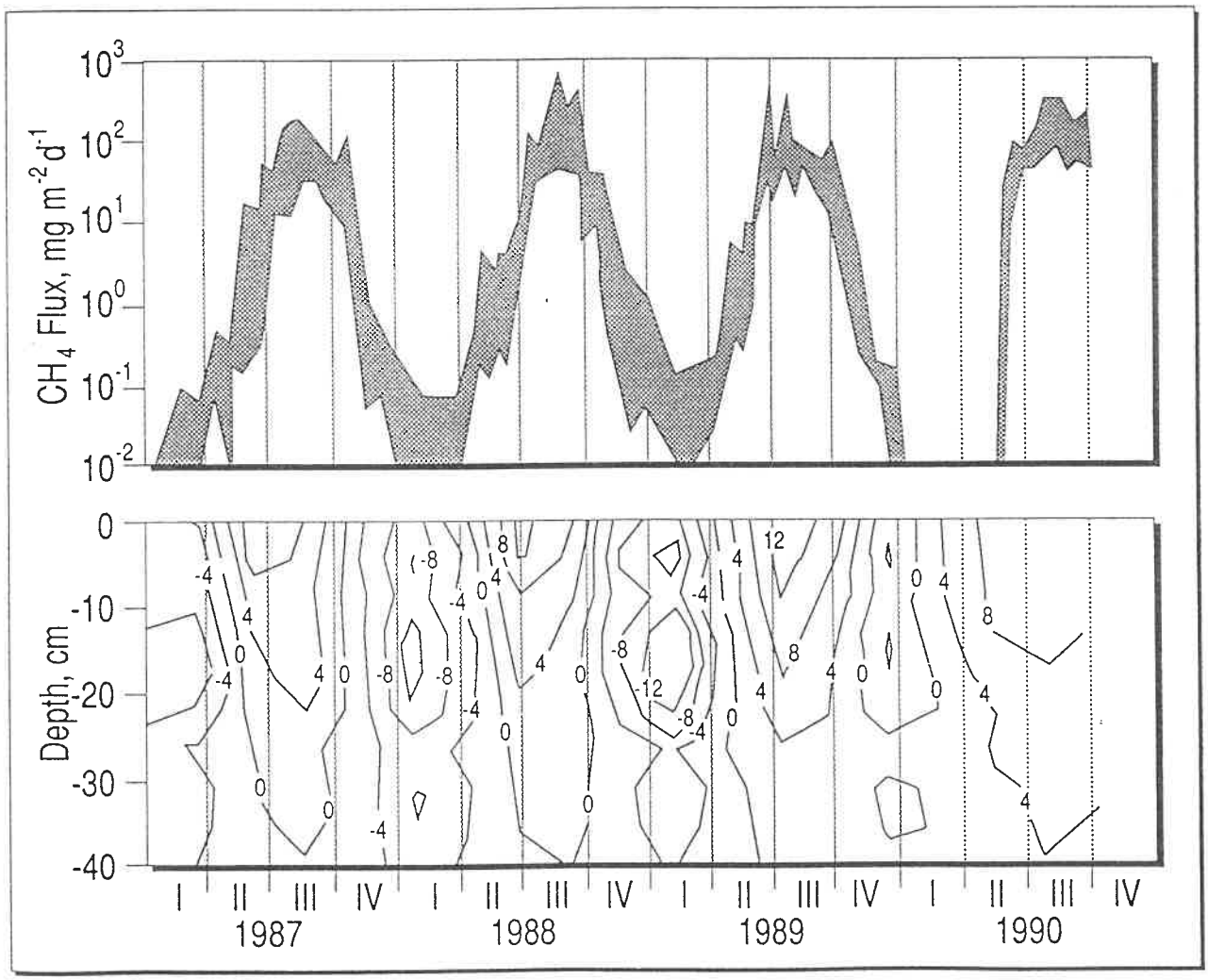


Figure 5.5. Seasonality in methane emission from tussock environments near Fairbanks, Alaska. Top: range of net CH_4 flux to the atmosphere. Bottom: soil isotherms (4°C contour interval) on depth versus time. Simplified from Whalen and Reeburgh (1992).

5.5.2. Moisture

Soil moisture has a fundamental control on methane emission because it is the natural precursor for anaerobic, methane-producing environments to develop. However, the amount of aerobic space between the water table and the soil surface also limits the potential for aerobic microbial methane consumption. Since in permafrost soils there is a vertical lower limit for production (the permafrost horizon), the water table applies constraints on both the production and consumption processes (Figure 3.2).

At moist sites, there is a relatively good correlation between the fluctuation of the water table and net emission with higher fluxes when the water table is close to the soil surface. Whether this is due to constraints on production or consumption, or a combination of both, is not entirely clear. However, Bartlett *et al.* (1992) and the present study show temporal correlations between the constantly falling water table at *Carex* sites and decreasing flux. A strong spatial control was illustrated in Figure 5.1 above. An inter-site linear correlation between water table and flux was found by Sebacher *et al.* (1986) and, in general, the importance of spatial differences in moisture content for net methane flux has been shown in many studies (Svensson and Roswall, 1984; Moore *et al.*, 1990; Bartlett *et al.*, 1992).

In July 1993 I carried out a preliminary study of methane flux at Stordalen, Abisko, Northern Sweden. I established a transect of 20 chambers at approximately 2 meter intervals starting from a hummock and crossing a waterlogged depression, ending on a second hummock. The hypothesis was that due to the moisture gradient, flux would be higher at the intermediate sites. Figure 5.6 shows how this expectation was fulfilled, but with some notable exceptions that underline the often chaotic picture arising when methane flux is correlated with a single variable.

Across the tundra as a whole, soil moisture might spatially be a good single indicator of the scale of net methane emission. However, temporal fluctuations form a more complex picture where no single parameter seems to be sufficient for predicting fluxes. Combined soil moisture, thaw depth and soil temperature are probably the best indicators of temporal fluctuations in net methane flux.

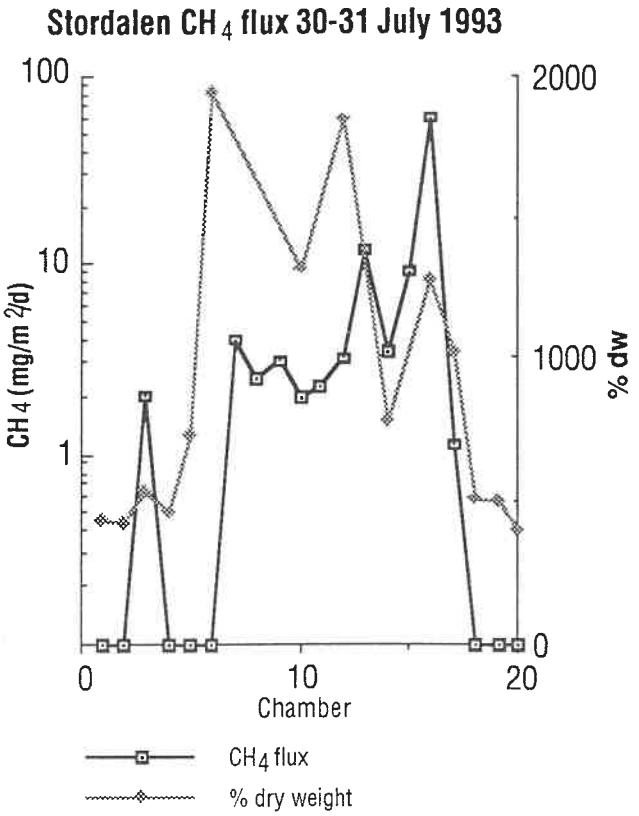


Figure 5.6. Methane flux and soil moisture content (% dry weight of top 10 cm soil) at 20 stations at approximately two meter intervals in Stordalen mire near Abisko, Northern Sweden. See Appendix 1 for details on methods for sampling and analyses.

5.5.3. Acidity and nutrients

Northern wetlands can be broadly separated into two major categories determined by hydrology and nutrient status: ombrotrophic bog and minerotrophic fen communities. Ombrotrophic bogs are "closed" wetlands which have precipitation and atmospheric deposition as the only sources of minerals and nutrients. Minerotrophic fens have a varying degree of water-flow through the system, which supplies minerals and nutrients. The consequence is a greater nutrient limitation and generally lower pH in the ombrotrophic systems compared with the minerotrophic.

The effect of this difference on net methane emission was shown by Svensson (1983). Minerotrophic mire areas had a significantly higher flux than ombrotrophic areas. A similar difference was found by Moore and Knowles (1989) in laboratory

experiments with fen and bog soil cores. The ratio of carbon dioxide to methane release at a range of moisture conditions in both studies was much lower in minerotrophic fen soils than in the ombrotrophic bog communities. This seems to indicate that neither carbon dioxide nor methane evolution is inhibited to any major extent by low nutrient status, but rather that methanogenesis is limited by lower pH in the ombrotrophic systems. I attempted to test this hypothesis in situ by adding lime to some experimental plots with low pH on the North Slope of Alaska. This experiment showed, however, no significant changes compared with control plots. It was a preliminary experiment and the inconclusive result might be due to a flaw in the experimental design (lacking means of insuring the effect of the lime actually propagated to the depth of methane production). It should be noted that not only production but also consumption of methane is influenced by pH (Dunfield *et al.*, 1993), the relationship is therefore complex in areas where oxidation plays a major role in determining net methane emission.

Nutrient availability, nitrogen in particular, is also linked to methane cycling in soils. Apart from the direct effects on methane oxidation mentioned in Chapter 3, soil fertilisation may also lead to a more dense vascular plant cover. The possible implications of this will be discused below.

The hydrology, nutrient, and pH characteristics of wetlands probably exert a major spatial control on regional differences in net methane emission. If climate change causes changes in the hydrology and nutrient availability of northern wetlands, this might also prove a long term temporal control on net emission from tundra environments.

5.5.4. Organic material.

Methane is produced in two different ways by methanogens: carbon dioxide reduction ahd acetate fermentation. Svensson (1983) showed how in cold, acid, environments the prevailing process is acetate fermentation. The isotopical analyses presented in section 5.3 seemed to confirm this. To test the possible substrate limitation of methane

production in tundra soils, I added acetate to two experimental plots in situ in a wet tundra environment . Figure 5.7 shows an episodic event of very high methane emission observed after acetate additions. Although the effect was not reproducible in a second chamber with the same treatment, the episodic event might still have been provoked by the sudden substantial increase in soil concentration of acetate. This provides a possible cause for the episodic events described and discussed in section 4.2.1 which are of great importance for the total tundra emission figure. "Hot spots" of extremely high microbial activity linked to sudden changes in substrate status are well known to occur naturally (Parkin, 1987).

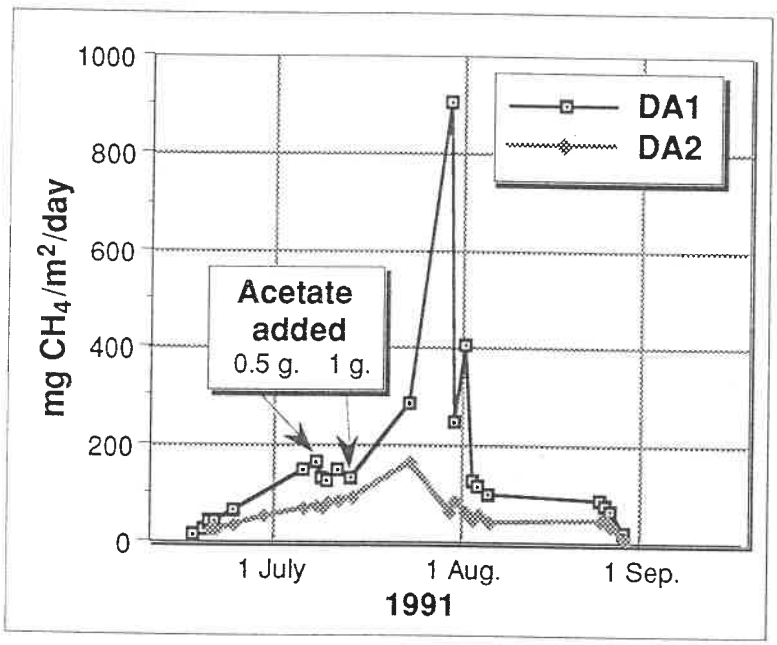


Figure 5.7. Net emission of methane from two experimental plots (DA1 & DA2) on the North Slope of Alaska to which acetate ($NaC_2H_3O_2$ dissolved in standing water) was added.

The quality of organic material and resulting substrate availability for microbial processes might cause moderate spatial differences in methane flux. If change in substrate level is in fact the trigger for episodic events, then a most important effect from the organic material will be in determining the frequency and scale of these events. However, it is hard to see how this will provide a major changing factor with

respect to the effects of climate change. In terms of the latter the most important control related to organic material is probably the increasing availability of organic material which will follow a deeper active layer caused by climate warming. However, quantifying this effect is extremely difficult due to the different decay potential with depth dicussed in Chapter 3.

5.5.5. Vascular plants and transport mechanisms

As shown in Chapter 4 and also by other studies (Whalen and Reeburgh, 1988; Bartlett *et al.*, 1992; Whalen and Reeburgh, 1992; Whiting and Chanton, 1992) vascular plant communities have significantly higher emission than non-vascular communities. Investigating the association between plants and net methane emission in a sub-arctic fen, Whiting and Chanton (1992) found that over 90% of the total emission was plant dependent transport. An implication of this association is that the plants provide a conduit which bypasses potential oxidation at the soil-water/atmosphere interface. Whiting and Chanton (1992) found a clear association between plant production, soil methane pools and emission, and that vascular plant biomass had a significant correlation with methane flux. I found similarly that the number of vascular plant tillers in moss pads showed a strong correlation with methane emission, and also that the removal of vascular plants at a wet site was followed by a dramatic reduction in the emission (Figure 5.8).

Vascular plants have therefore an important spatial control on methane emission from tundra soils. This might have long-term temporal consequences for net emission from tundra areas provided changing climate somehow favoured vascular compared to non-vascular plant communities. The question about how climate change will affect soil nutrient status is obviously also linked to this.

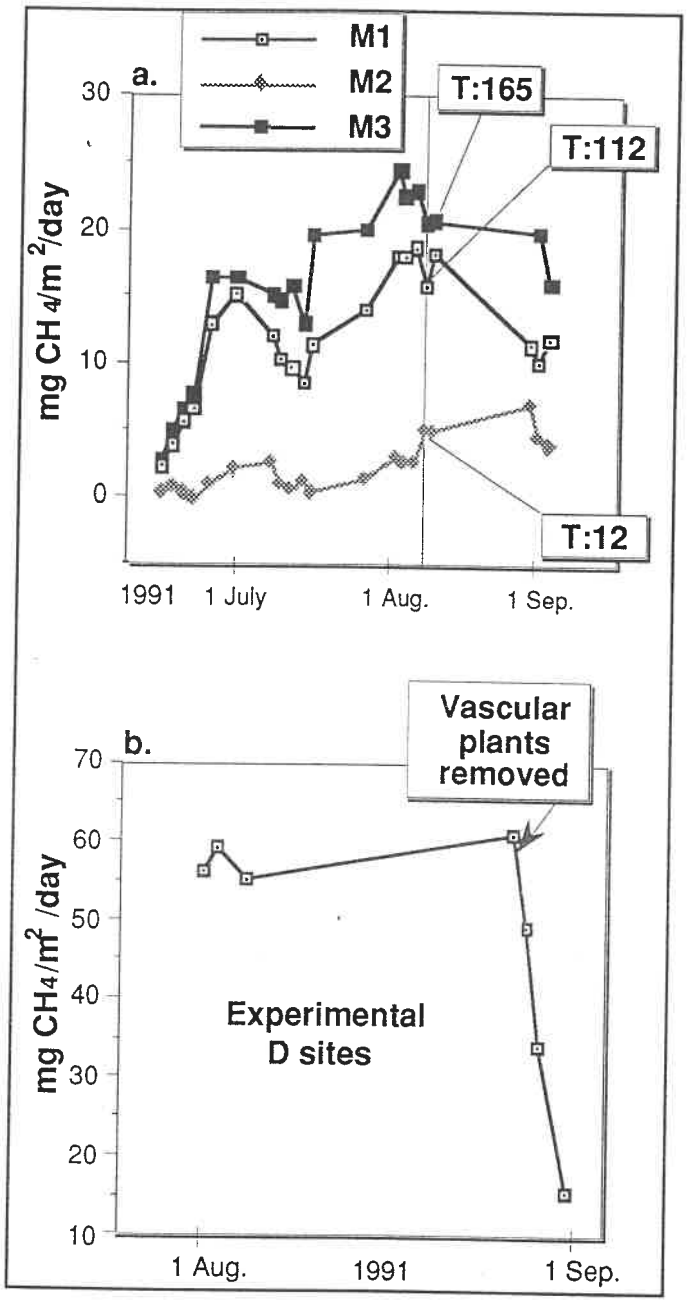


Figure 5.8. The effect of vascular plants on net methane emission from tundra environments: **a**. Net emission from three moss pads. The number of vascular plant tillers (T) in each moss pad was counted on 5 August 1991 and showed strong correlation with net emission (r^2=0.99). **b**. Mean of net emission from four wet tundra sites (D sites) before and after removal of vascular plants.

Methane which is not oxidised or transported to the atmosphere by diffusive processes is occasionally released as bubbles or "ebulliative" flux. The relative amount of methane released by ebullition is extremely difficult to assess. It has been estimated that as much as 85% of the methane released from tropical swamps are bubbles (Devol *et al.*, 1988). In tundra environments this percentage is probably much lower (Reeburgh *et al.*, 1993), but quantifications of the ratio are rare. In a wet tundra/small lake environment in Alaska, Martens *et al.* (1992) found between 0.6% and 17% of the regional flux to be bubbles.

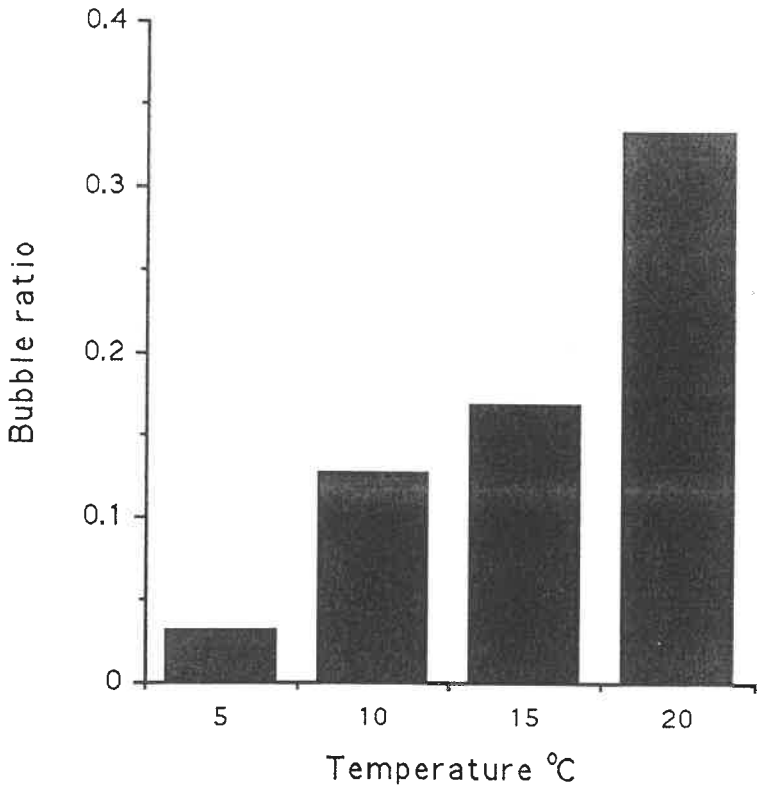


Figure 5.9. Ratio of core flux meaurements where the time-series of samples was disturbed by release of bubbles, to the total numbers of measurements at different temperatures. Total numbers of measurements were: 32 at 5°C, 16 at 10°C, 24 at 15°C and 8 at 20°C.

In the soil core experiments outlined in section 5.3, bubble methane was observed at a higher rate as the temperatures increased (Figure 5.9). This feature seemed more associated with the immediate temperature than with the point in the progressing "season", which indicates a temperature dependent physical process as cause for the bubble release. This crude observation seems to fit the broad picture of higher ratios of bubble methane being released at warmer latitudes.

5.5.6. Plant production

Encouragement for satelite-based extrapolations of global wetland CH_4 emission was provided by Whiting and Chanton's (1993) recent finding that emissions at a variety of North American wetland and acricultural rice sites were linearly correlated with net ecosystem production (NEP). This correlation is linked to the relationships described in the former section on vascular plant control on emission. At constantly wet sites, the denser the vascular vegetation, the higher the flux.

The relationships found by Whiting and Chanton apply, however, only to permanently inundated soils with vascular plants. If, for example, the tundra subdivision production in Figure 3.1 were correlated with mean emission, an almost negative correlation between flux and production would be found (shrub tundra has the highest NPP and lowest flux; wet tundra has low NPP but highest flux).

5.5.7. Light intensity

King (1990) provided evidence that light induced changes in oxygen availability in the upper layers of a wetland soil (caused by controlling algae photosynthetic activity) produced an important control on microbial methane oxidation and, hence, net emission from the soils. However, neither I (see section 7.3) nor Whiting and Chanton (1992) found significant differences in methane emission based on light and dark chamber measurements. The latter study showed, furthermore, relatively small diurnal changes in net emission to be due to temperature changes rather than variations in light intensity. I compared diurnal measurements in mid summer (24 hour sunlight) and late

season (several dark hours), also finding no evidence of an effect due to changing light regime.

Light seems therefore not an important control on net methane emission. Even if it were, it would probably be of more importance for methodological discussions (dark versus light chamber techniques) than for climate change related analysis of controls on net emission.

5.6. Summary

No single parameter relationship between one environmental factor and CH_4 flux covering all sites in the field work was obtained. However, inter-season variations in CH_4 flux at dry sites seemed to be largely controlled by water table fluctuations while wetter sites were controlled by soil temperature. An increasing emission with the progressing season was observed and quantified as a function of degree days. This increasing "background" emission was shown in the laboratory as partly being caused by release from increasing amounts of dissolved methane in the pore water. When the wet tundra flux observed in the field was corrected by a factor dependent on degree days, the flux showed significantly better correlation with soil temperature.

Table 5.5 shows an attempt at spatial and temporal scales to summarise an estimated relative influence of the factors controlling methane flux based on the discussion above. In a climate change context ("long-term temporal") soil temperature, moisture and thaw depth are estimated to be the factors of highest importance. Hence, when modelling, an integration of the effects of soil moisture, thaw depth and soil temperature is probably the most immediate indicator of net methane flux response to climatic change. Also experimental evidence from boreal peatlands support this view. Here water table and temperature were recently shown as potential tools for predicting methane flux (Dise *et al.*, 1993).

It appears also that each of the production and oxidation processes have their own independent relationships with temperature and moisture. The balance between the impact of these parameters on the two processes gives the net emission from the soil. It

should be noted, however, that the two factors are interlinked also in terms of the thermodynamics of the soil thermal regime (moisture is an important control on the soil thermal diffusivity) and in terms of their effects on different components of the microbial community. Hence, the two factors need coupling when establishing predictive models. Such models should include also parametrisation of the soil thermodynamics. An attempt to develop such a model will be described in the following chapter.

Table 5.5. Estimated relative influence of spatial and temporal controls on net methane emission from tundra environments. The grading of influence is: --, none; *, minor; **, moderate; ***, major. Dry and wet tundra are separated under soil temperature and moisture since the effects are assumed to differ on a seasonal scale.

	Spatial	Short term temporal	Long term temporal
Soil temperature			
dry tundra	*	**	***
wet tundra	*	***	***
Soil moisture			
dry tundra	***	***	***
wet tundra	***	**	***
Nutrients and pH	***	*	**
Organic material	**	*	**a
Vascular plants	***	*	**b
Light intensity	--	*	--

[a] depending on decay potential of extra organic material made available for decomposition following a possible deeper summer thaw

[b] depending on possible vegetation changes

Chapter 6

Modelling methane emission

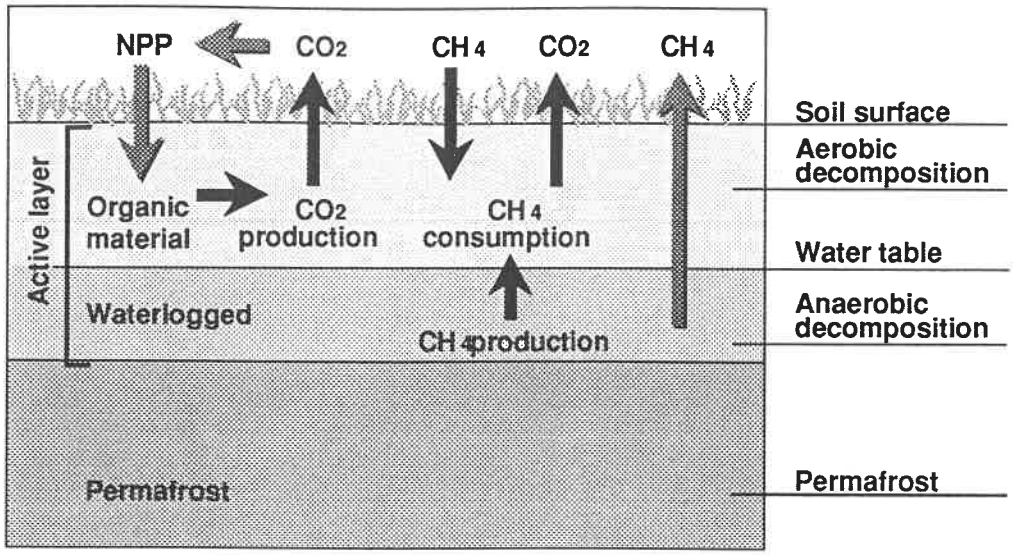
6.1. Introduction

There seems to be a general agreement among the current generation of global circulation models (GCM's) that (a) increasing concentrations of greenhouse gases in the atmosphere will cause a warmer climate globally and (b) that high northern latitudes will experience the greatest warming (Chapter 1). A general rough estimate for the region between 50 and 70°N is a warming in winter of 4°C and in summer 2°C (IPCC, 1992). When the 2 x CO_2 situation is expected to arise depends on complex issues such as development of the global energy policy and third world population growth. According to recent simulation models, doubling can be expected sometime around the middle of next century (see Chapter 1, IPCC, 1992).

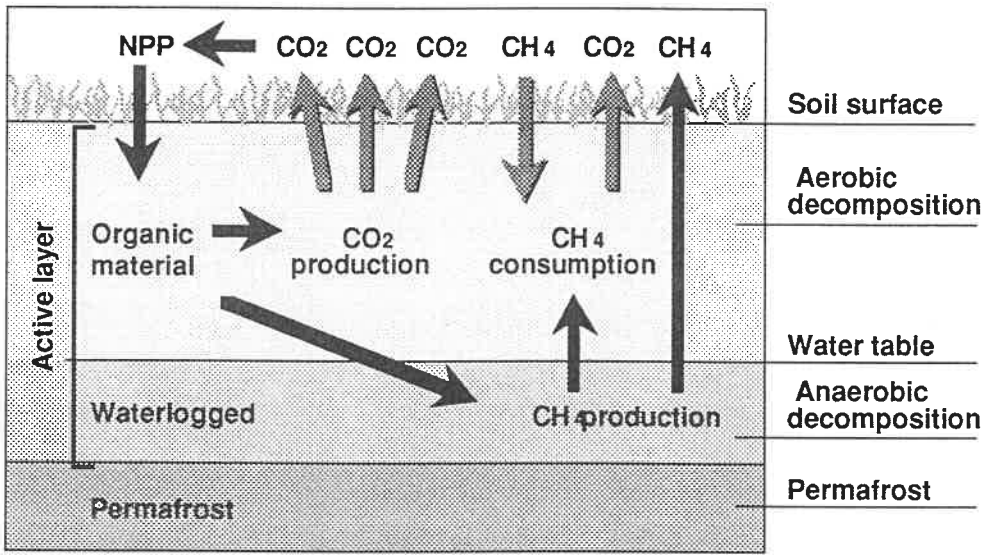
Predictions of precipitation changes are much less certain than temperature in the models. Thus, how the temperature/rainfall changes will translate into changes in soil moisture is even more uncertain. This is nevertheless an important question since the soil moisture change possess significant influence on how the trace gas balance of ecosystems will respond to climatic change. Such predictions are therefore of great importance when modelling further development of the greenhouse effect. Tundra, as shown in previous chapters, already being a substantial source of methane, could increase or decrease emissions providing feedback mechanisms of significant importance to climate models.

Qualitatively there is general agreement in the literature on the relationships between tundra trace gas flux and climate illustrated in Figure 6.1. A warmer wetter environment will probably increase methane emissions while a warmer and drier environment might decrease methane emission and possibly change the tundra from a carbon sink to a net source of CO_2. However, given the complexity of factors controlling methane emission from tundra environments shown in Chapter 5, it is not a

NORMAL TUNDRA SOIL



WARMER AND DRYER



WARMER AND WETTER

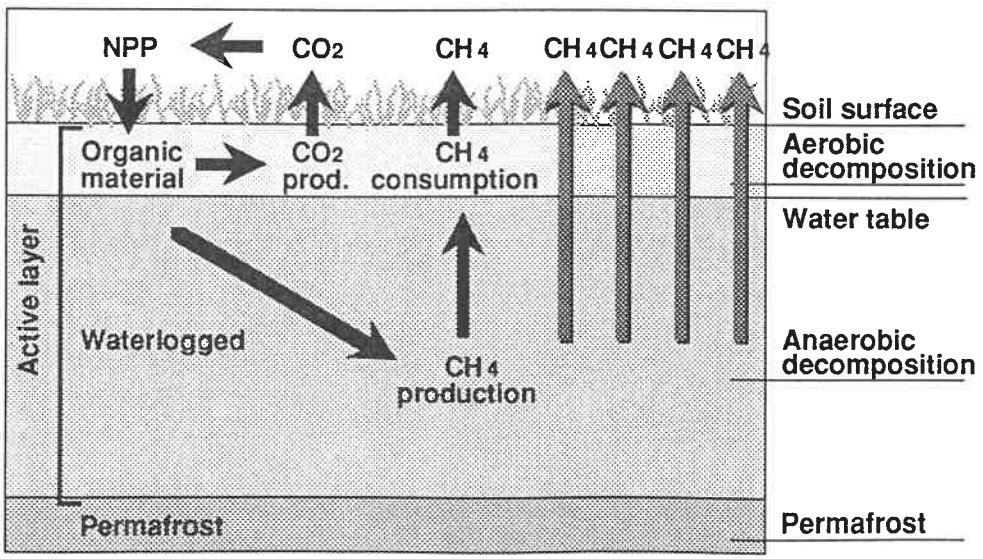


Figure 6.1. Preceding page. General hypothesis for the response of tundra exchange with the atmosphere of methane and carbon dioxide, following two scenarios of climate change.

simple task to quantify the relationships and integration of a number of parameters has been shown necessary. This chapter explores one attempt to improve a soil model developed at the Hadley Centre, Meteorological Office. The model and how we at SPRI and collaborators at the Hadley Centre have changed it during the course of the project is described. I will focus on my own part of the work, involving the methane routine, and discuss the results that have so far emerged from experimental runs of the model.

6.2. The Meteorological Office model

As described in Chapter 1 climate simulations are carried out with numerical General Circulation Models (GCM) which have been developed from weather forecast models. For investigations of climate change due to increased greenhouse gas concentrations, they have generally been run with simple representations of the upper ocean and recently with more detailed dynamical models of the ocean to its full depth. Relatively simple schemes for interactive land surface temperature and soil moisture are also included. Representations of other elements of the climate system like land-ice and biosphere have usually been included as non-interactive components. These representations are, in the Meteorological Office GCM, referred to as the "Met. Office land surface scheme". The scheme has recently undergone improvements by the inclusion of multi- rather than single-layer soil hydrology and also, as part of the present project, by the introduction of permafrost thermodynamics and soil water phase changes.

The Meteorological Office GCM consists of a large number of joined single column models (SCM) which are one-dimensional models of the vertical structure of the atmosphere and basically represents single gridboxes of the GCM. The SCMs joined and run as the GCM is termed the UK Meteorological Office Unified Model (UM) (Lean, 1991). In the SCM, the effects of large-scale horizontal and vertical

motion are either treated statistically or taken from observations, allowing the surface and atmosphere to be forced with realistic, time-varying atmospheric conditions.

In the present context SCM is as far as we to date have taken the permafrost and methane model we have developed. Full GCM runs are obviously very expensive in computer time and it is also of little use trying out new, poorly-tested parametrizations on a GCM. It is, however, the plan that the new soil and methane routines will enter the UM early in 1994.

6.3. Hydrology and thermodynamics

The model's hydrology and thermodynamics are based on vertically discretised forms of Richards' equation of fluid flow in a porous medium and Fick's law of heat diffusion. Soil-water phase changes are parametrised in terms of a simple dependence of maximum unfrozen water on temperature (Williams and Smith, 1989), thereby simulating the observed phenomenon of unfrozen water existing at temperatures below 0°C (Williams and Smith, 1989). Latent heat effects are included through an effective (temperature dependent) heat capacity. The hydraulic conductivity and soil water suction are parametrised in terms of the unfrozen soil water only (Black and Allen, 1988). In this way the simulated frozen soils exhibit the observed properties of low hydraulic conductivity and strong water suction. The model can be used with an arbitrary number of soil layers, but the results below were produced using four soil layers with thicknesses (from the surface downwards) of 3.8 cm, 14.5 cm, 48.5 cm and 168.8 cm.

The total soil depth of, in this case, 2.36 meter is not realistic in most tundra environments. The relatively deep bottom layer is, however, necessary to correctly represent the heat capacity of permafrost.

6.4. Methane model

6.4.1. Controlling factors chosen

Following the conclusions in the preceding chapter, soil temperature, soil moisture and thaw depth are considered the most important factors controlling temporal variations in methane emission. In order to incorporate and integrate these effects in the model a simple mechanistic approach was taken.

Production and consumption was assumed to have the same standard microbial "Q_{10}" temperature dependencies. This is inaccurate from a microbiological point of view since most studies of methanogens show higher Q_{10} dependencies than those of methanotrophs. However, large variations in Q_{10} values have been found and no specific studies of the microbial populations concerned are available from the Toolik area where the emission data used as validation was obtained. Also, rather than attempting to produce a specific microbiological model for a particular site, the aim of this work was to make a broader ecosystem-type model. The most simple and general temperature dependencies mentioned above were therefore adopted.

Fractional saturation was used as the moisture constraint on methane production and consumption. The maximum thaw depth was considered a lower limit for microbial activity in the soil, despite the possible limited sub-zero activity described in Chapter 3. All unfrozen soil layers were given the same potential for decay, which is probably unrealistic considering the different decay potentials with depth discussed in Chapter 3. The root zone is in the model for the purpose of parametrising evapotranspiration and it may in the future be introduced as a zone of higher decay potential ("productivity") compared with deeper layers. There are preliminary results suggesting most methane production is associated with the root zone (J. Schimel, pers. comm., 1993). A general picture showing the factors influencing methane dynamics in the model is found in Figure 6.2.

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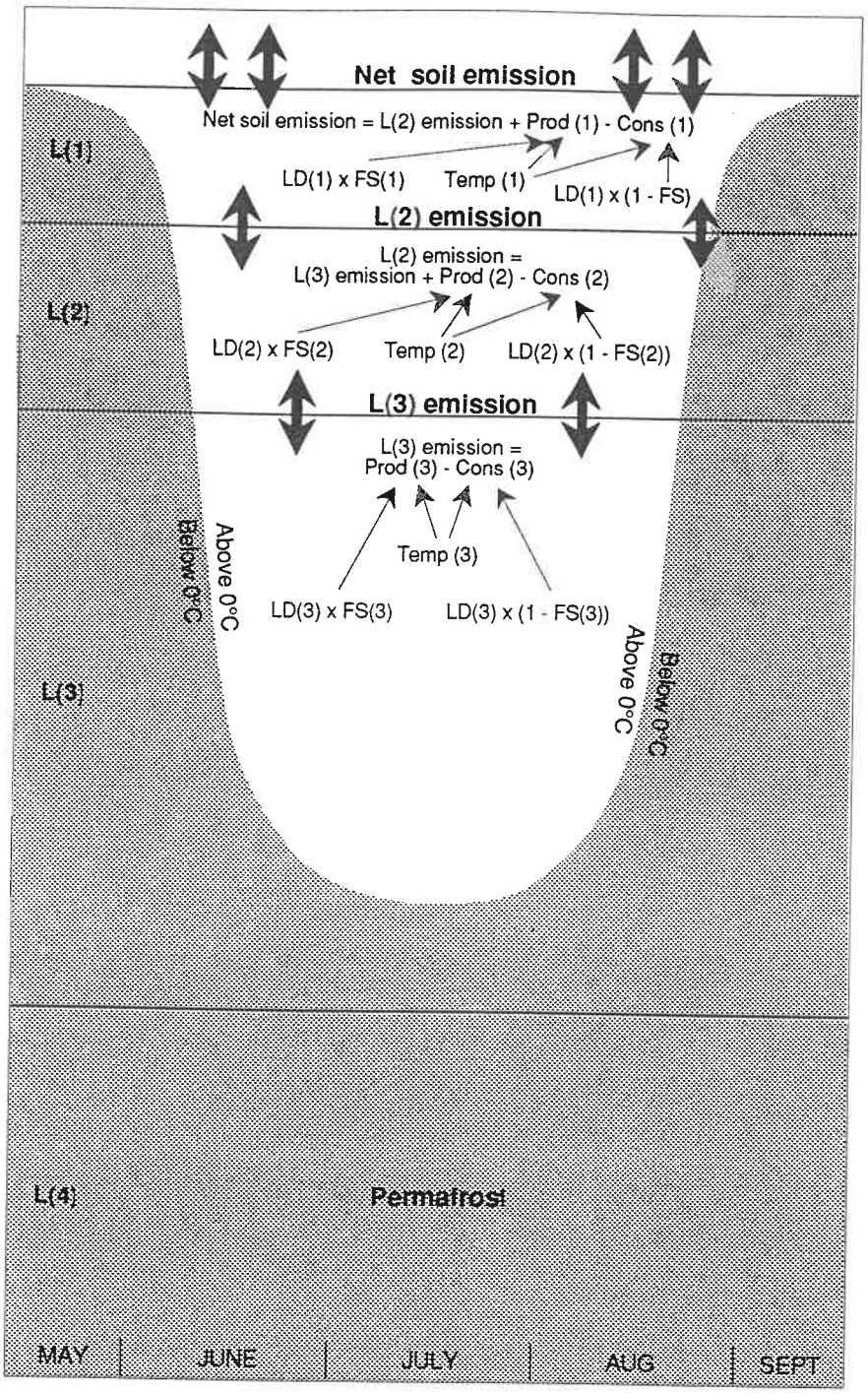

Figure 6.2. Simplified schematic illustration of factors influencing seasonal methane production and consumption in the model (arrows indicate influence). The depths of the soil layers are not to scale but indicative. FS is fractional saturation derived from the unfrozen water content and LD is layer depth. Figures in brackets are layer numbers.

6.4.2. Parametrisation of methane emission

The parametrisation developed for this study is based on the assumption that each soil layer can behave as a methane-producing or methane-consuming unit depending on its temperature and moisture content. Layers which contain more than half the saturation soil moisture are treated as net emitters, whilst those with less than half are net consumers. As mentioned above, the microbial temperature dependence is the commonly used "Q_{10}" for layers which are unfrozen, while partially-frozen layers ($T <$ 0°C) are assumed to be passive (zero net emission). Thus the model of net soil methane flux, F_{CH4}, takes the simple form:

$$F_{CH4} = \sum_{i=1}^{N} f(T_i)\,\{2\Theta_i - 1\}\,(k_i \Delta z_i) Q_{10}{}^{(T_i - 2)/10}$$

where Θ_i is the fractional saturation of the ith soil layer, T_i is the temperature in°C of the ith soil layer, N is the total number of soil layers (4 in this case), and$f(T_i)$ is the step function taking the value 1 for $T_i > 0$°C and 0 for $T_i < 0$°C. k_i and Δz_i are the methane productivity and depth of the ith layer respectively. In the simulations presented here $Q_{10} = 2.0$. Note that decomposition is allowed potentially (depending on temperature) to occur at the same rate in all layers which have temperatures above 0°C. As discussed above this is probably unrealistic (due to the varying decay potential with depth) and in any case the model is only strictly applicable to areas where the organic layer is deeper than the maximum thaw depth (i.e. organic wet/moist tundra).

As shown in Chapter 5, the methane emission increases during the thaw season as a function of degree days. Similarly, in this model the layer productivity, k_i, has a simple linear dependence on degree days:

$$k_i(\tau) = 50\left\{\alpha_i + \beta i \int_0^{\tau} T_i dt\right\}$$

where α_i = 0.85 mg CH_4 m^{-2} day^{-1} and β_i = 0.00075 mg CH_4 m^{-2} $day^{-1}°C^{-1}$.The layer productivity is set at 50 mg/m^2/d based on mean emission in the observations under "standard" conditions. Again, this productivity should ideally vary with depth. Also the similar conditions given to production and consumption could theoretically under extremely dry conditions give rise to unrealistic high rates of consumption. In order to avoid this, account would have to be taken of physical limitations on gas transport in the soil.

6.5. Results

The following experiments have been carried out with the model:

1. a local simulation of variations in methane flux between 15 June and 5 August 1991 using hourly weather observations of air temperature, precipitation, net radiation and wind speed obtained by the Long Term Ecological Research (LTER) Programme near Toolik Lake on the North Slope of Alaska (Figure 6.3). Results were compared with in situ observations of methane flux over the same period near Toolik Lake (Chapter 4). Unfortunatly the length of the period of this validation run was constrained by the availability of consistent LTER weather data.
2. a number of sensitivity experiments where the LTER weather data were manipulated with respect to input parameters air temperature and rainfall, creating warmer, colder, wetter and drier summers in various combinations.
3. a regional 5-year simulation where the Meteorological Office single column model (Dolman and Gregory, 1992) was forced with climatology for the North Slope of Alaska (68°N, 149°W) derived from Meteorological Office operational analyses (R.E. Essery, pers. comm., 1993).
4. as 3 but for 2 x CO_2 with an assumed warming in the climatological forcing.

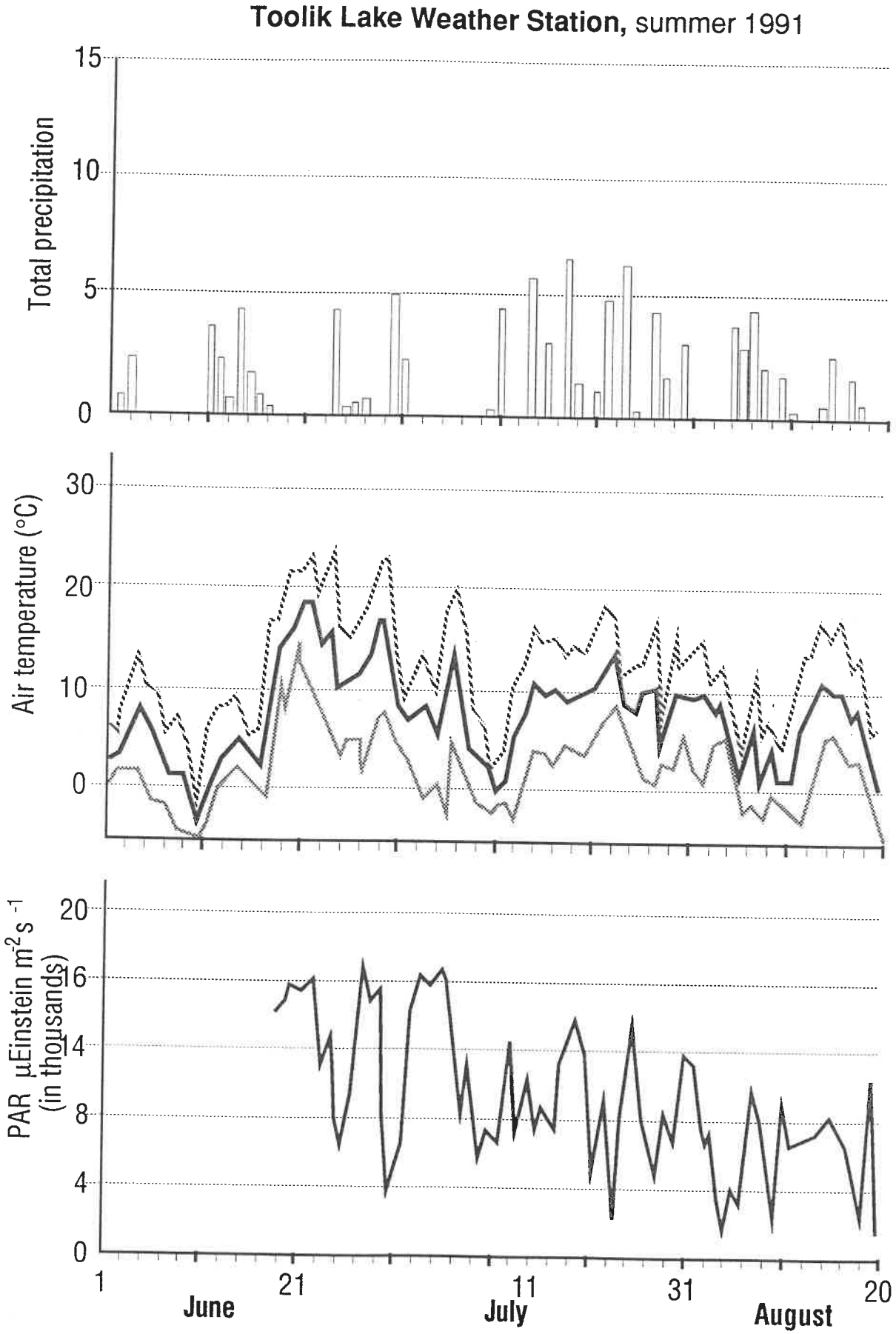
Toolik Lake Weather Station, summer 1991
Total precipitation
15
10
5
0
Air temperature (°C)
30
20
10
0
PAR µEinstein m-2 s-1
(in thousands)
20
16
12
8
4
0
1
21
11
31
20
June
July
August

Figure 6.3 (preceding page). Weather data from the summer of 1991 at Toolik Lake on the North Slope of Alaska. Data from 15 June to 8 August are input to the model experiment described.

6.5.1. Stand-alone tests with LTER data

Figure 6.3 shows part of the weather input to the stand-alone version of the model and Figure 6.4 shows a comparison between measured and simulated methane fluxes from model experiment 1. The measured soil environment (soil temperature, water table and thaw depth) was reproduced to within an error of about 10% by the model. The field sites used for the comparison are the E sites because all three stations here behaved similarly and in relative accordance with an overall mean flux. The E sites also represent an intermediate environment between wet and moist tundra.

Figure 6.4 shows a reasonable agreement between modelled and real data with a slightly more extreme variation in the observations. Most of the model run falls within the standard error of the observations. A linear regression analysis of the two data sets based on the days of observations show $r^2 = 0.75$ ($n = 15$). The mean values are very similar, 24.5 and 25.7 mg CH_4 m^{-2} day^{-1} for the observed and modelled data respectively.The model run presented in Figure 6.4 provided some confidence that despite the simple approach taken the model does seem to integrate the main controlling factors in a realistic way.

The stand-alone version of the model was then manipulated with respect to air temperature and precipitation. Air temperature was increased and decreased by 4°C. Versions with unchanged and warmer air temperatures were combined with changes in precipitation varying from a 50% decrease to a 50% increase in rainfall. Details of input and output parameters from the sensitivity runs can be found in Appendix 3.

The thaw depth increased by 11% following a 4°C increase in air temperature and this effect varied little with different precipitation changes. A 4°C decrease in air temperature caused a 23% shallower active layer and periodical surface freezing during the cold period in early July. The fluctuating water table was directly correlated with the varying precipitation and, since the model did not produce significant variations in

the thaw depth as a consequence of precipitation changes, this caused methane emission to be linearly correlated with fractional precipitation (Figure 6.5). Figure 6.5 shows the mean methane emission in two temperature scenarios (0 and + 4°C changes) as a function of fractional precipitation.

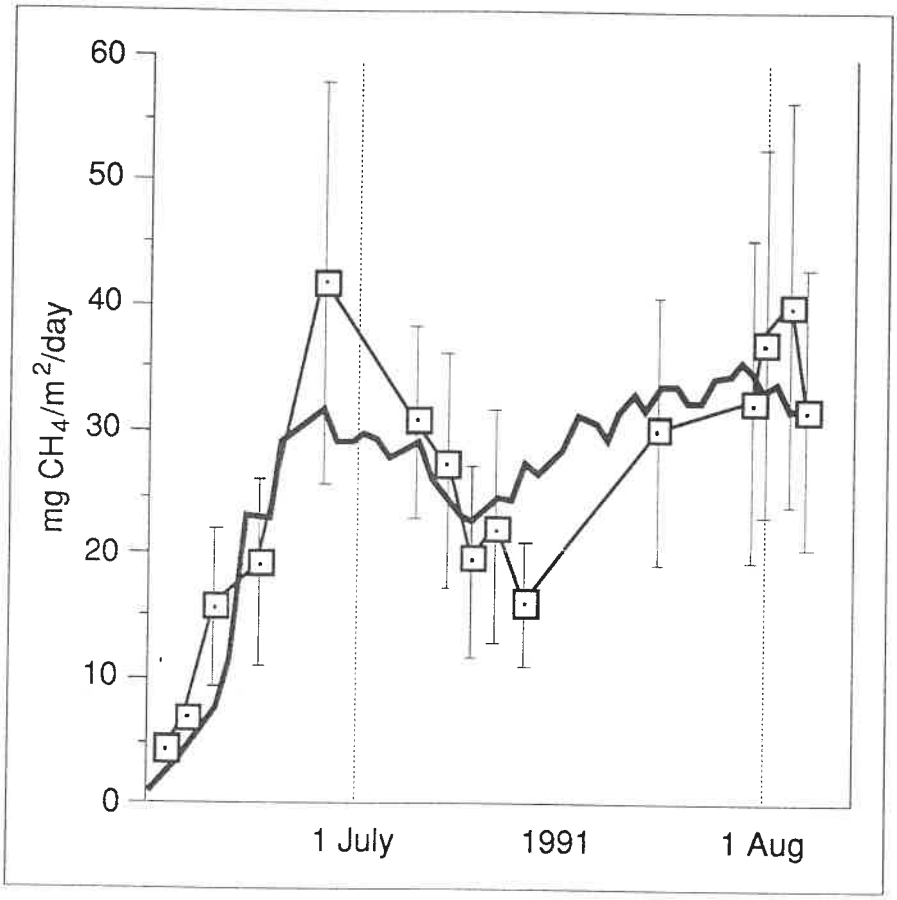

Figure 6.4. Methane flux from Arctic tundra near Toolik Lake on the North Slope of Alaska between 15 June and 5 August 1991 as measured in the field and simulated by the model (bold line). The bars indicate the standard error in the mean of the observations.

Figure 6.5 illustrates how the stand-alone version of the model predicts a 13% increase in methane emission with a 4°C temperature increase and no change in precipitation. Correspondingly, approximately 13% decrease in precipitation is needed for drying to outweigh the effect of air warming and result in zero change in net emission. The model seems slightly more sensitive to precipitation changes under the

warming scenario. With a 50% reduction in rainfall there is little difference in flux between the two temperature scenarios while at a 50% increase in precipitation warming produces about 17% higher flux. According to this version of the model a mean 4°C increase and 10% increase in precipitation would cause a 21% increase in methane emission. A similar warming but with a 10% decrease in precipitation would result in a 5% increase in emission.

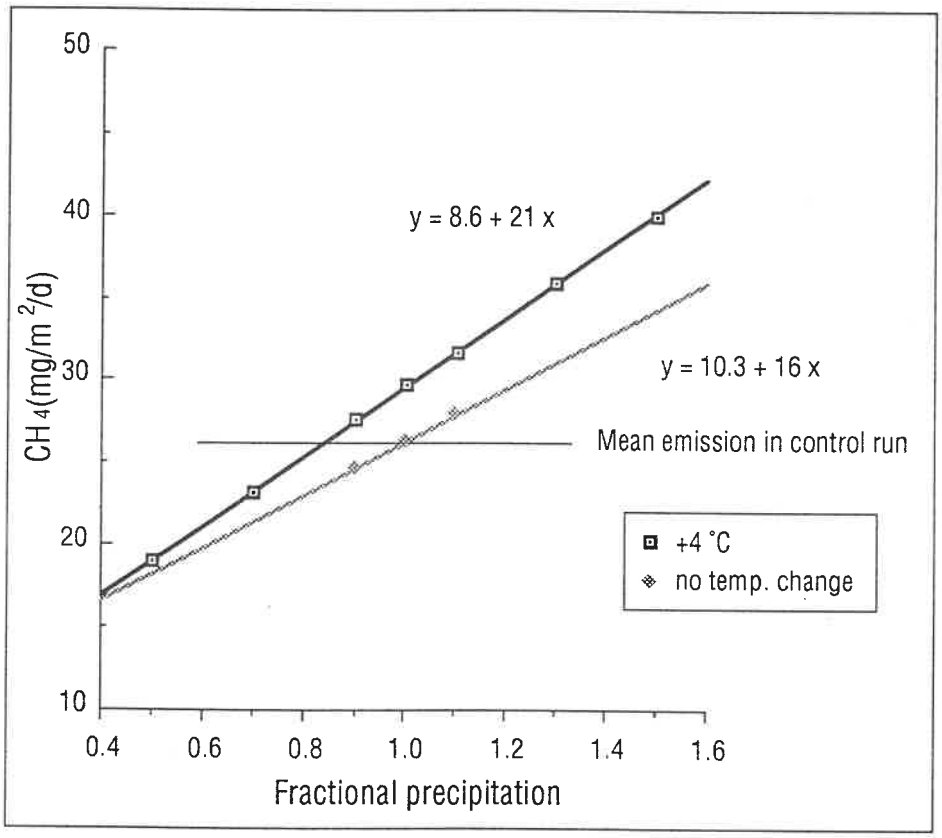

Figure 6.5. Mean methane emisssion output from the stand-alone model driven with manipulated LTER weather data (see text). A fractional precipitation of 1 represents the observed rainfall, 1.3, for example, a 30% increase.

6.5.2. Tests with operational analysis

Selected model variables from experiments 3 and 4 are shown in Figure 6.6 and details given in Appendix 3. It should be noted here that the thaw depth, active layer temperature and active layer soil moisture were not explicitly modelled, but diagnosed from the layer temperatures and moisture contents. Details are given in the figure

caption. In experiment 3 (1 x CO_2) the thaw depth reaches approximately 1 metre which is slightly deeper than what is normally found in tundra environments. The annual mean methane flux of 17 mg CH_4 m^{-2} day^{-1} is within range of most recent field studies of tundra methane emission including the present (see Chapter 4). The significant interannual variability in these experiments is produced by the random forcing of the single column model using the variances of temperature and dewpoint depression derived from the operational analyses (Dolman and Gregory, 1992; R.E. Essery, pers. comm., 1993). Years two and five (Figure 6.6) show a relatively large mid-summer drying (i.e. lower active layer soil moisture) and a correspondingly reduced methane emission.

Finally a 2 x CO_2 experiment was carried out. The mean air temperature used to force the single column model was increased by 4°C in the winter and 2°C in the summer (IPCC, 1990; IPCC, 1992). The resulting five year simulation is also shown in Figure 6.6. Table 6.1 compares the five year means from model experiments 3 and 4. The increase in precipitation is largely due to the simple warming scenario assumed, i.e. a warming of the climatological temperature profile but no change in the profile of dew point depression. However, the resulting precipitation change is within the range of GCM predictions (IPCC, 1990; IPCC, 1992). The most striking difference occurs in the mean thaw depth which increases by 42% in the 2 x CO_2 experiment. This reflects both a deeper maximum thaw and a longer thaw season (Figure 6.6). The latter is only partly responsible for the increases in active layer soil moisture and temperature. The combination of slightly warmer and moister soils with increased active soil volume leads to a 56% enhancement of methane emission in the 2 x CO_2 experiment.

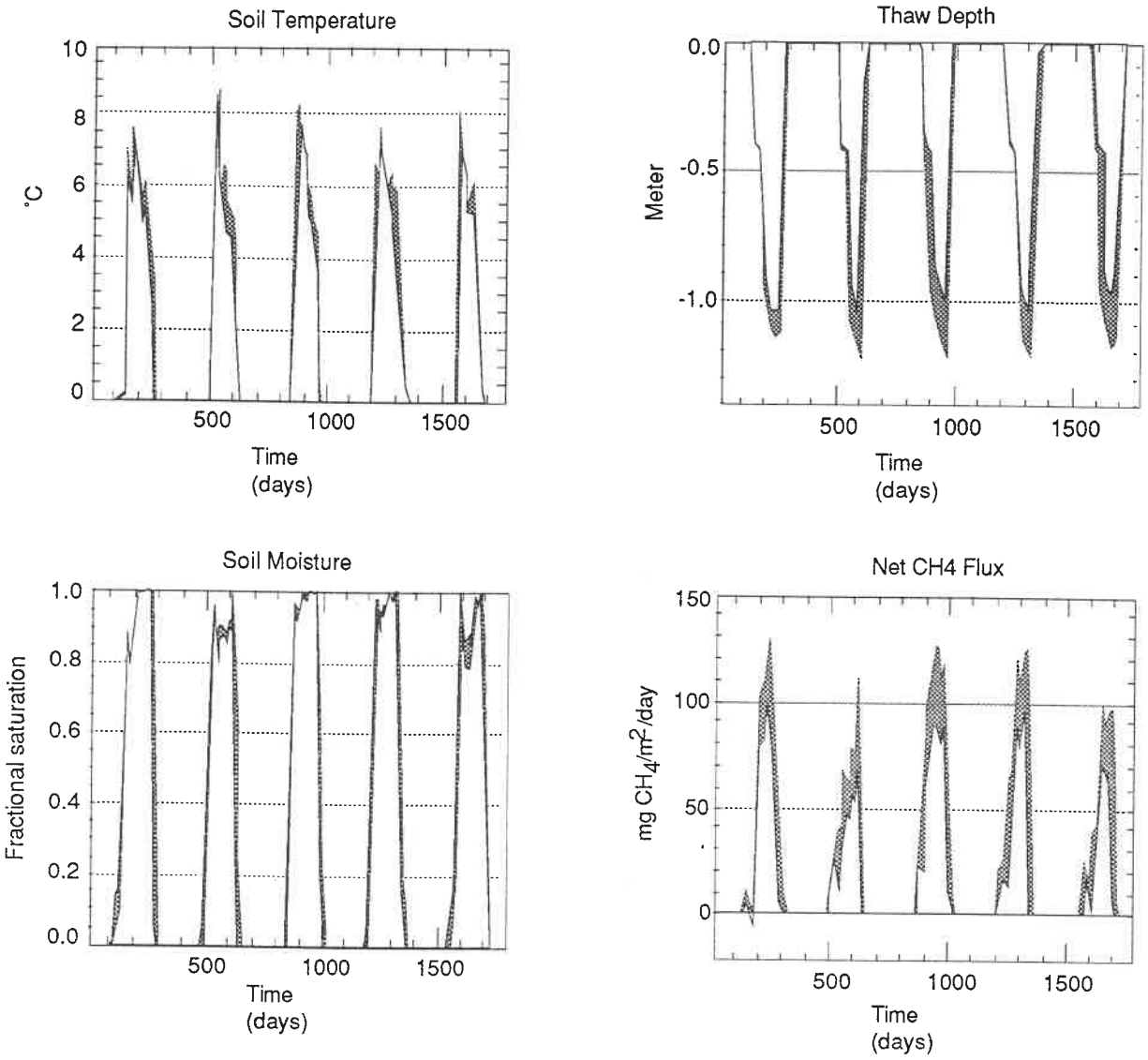

Figure 6.6. Active layer temperature, thaw depth, active layer soil moisture (as a fraction of saturation) and net CH_4 flux from five-year model runs using forcing data from the Meteorological Office operational analyses. Results from both the 2 x CO_2 and control (1 x CO_2) experiments are shown, with the black lines encompassing white areas representing the control simulation and the stippled areas the increases which occur in the 2 x CO_2 experiment. The thaw depth is defined as the depth of the 0°C isotherm, whilst the soil temperature and soil moisture represent the (vertically integrated) mean values for the active layer, i.e. the layer from the surface to the 0°C level. All three take the value zero if the 0°C isotherm reaches the surface.

Table 6.1. Five year means from the 1 x CO_2 and 2 x CO_2 simulations. The active layer soil moisture is expressed as a fraction of saturation and is the annual average, which means it includes the winter where very little unfrozen soil moisture is present.

	1 x CO_2	2 x CO_2	2 x CO_2 - 1 x CO_2
Precipitation (mm day^{-1})	1.63	1.93	0.30 (+18%)
Surface temperature (°C)	-11.1	-7.1	4.0
Active layer temperature (°C)	1.55	1.87	0.32 (+21%)
Thaw depth (cm)	22.5	31.9	9.4 (+42%)
Active layer soil moisture	0.298	0.349	0.051 (+17%)
Methane emission (mg CH_4 m^{-2} day^{-1})	17.1	26.6	9.5 (+56%)

6.6. Discussion: predictions

There are very few published models attempting to predict tundra methane emission response to climate change available for comparison. Roulet *et al.* (1992) used simple and separate hydrological and thermodynamical models for floating and non-floating northern fens to arrive at estimates for soil temperature and water table response to a warmer climate. They linked the responses to methane flux by separate relationships, thus not integrating the effects. They estimated that temperature increase alone would increase fluxes by between 5 and 40% which is in accordance with the results presented in the sensitivity study above. Their model predicted a falling water table level following climate warming, and methane flux was highly sensitive to this, with a decrease in flux of up to 81% following a water table that dropped from 8 to 22 cm below the peat surface. We did not see such dramatic changes in water table in our scenarios and thus not such extreme changes in flux. This may be due to our model being adapted to a permafrost environment with no vertical drainage and limited surface runoff.

Harriss *et al.* (1993) related methane emission to past temperature records for various northern wetland sites in an attempt to show the sensitivity to climatic

anomalies. They mostly found rather modest interannual variations of 0-2 Tg/yr or 0-12% of the total model flux. They note how these variations are unlikely to have influenced changing atmospheric concentration of methane over the past century and that climatic change will have to produce uniform increases in soil moisture over all northern wetland regions if these are to produce any significant feedback effect on greenhouse warming. During initial stages of such warming Harriss *et al.* postulate that regional differences in changes of the soil moisture regime will outweigh any significant contribution to further warming.

The SCM runs carried out in this study are probably more reliable than the stand-alone sensitivity tests. The stand-alone runs showed relative sensitive to the initial conditions which in some cases were difficult to estimate. Also the SCM runs incorporate the effect of warming on the annual cycles including those of a longer thaw season. However, the SCM only produced one warming scenario which makes the sensitivity of the model difficult to compare with other models. The results of stand-alone version were therefore used in the following comparison. Similarly, in order to assess the effect of regional differences in climate warming the stand-alone version were used.

The sensitivity results presented here disagrees with the conclusions of Harriss *et al.* (1993) and Roulet *et al.* (1992) in that a warming and (modest) drying in our model still produces a higher output of methane. Say that the tundra at present emits 35 Tg CH_4/yr. If 50% of the tundra experiences warming and a 10% increase in precipitation, and the other half warming and a 10% decrease in precipitation, then according to our model, total flux would increase would be 17.5 x 13% = 20 Tg plus 17.5 x 5% = 18 Tg yeilding a total of 38 Tg. Although this is a sizeable increase in global tundra flux, 3 Tg/yr represents less than 1% of the total atmospheric input. This is clearly a very crude calculation that has no basis in actual predictions. However, it does seem to support the general conclusion drawn by Harriss *et al.* that a significant feedback on global warming from increased methane emission can only be expected if soil moisture increases uniformly across the tundra region.

There are obvious shortfalls in the model as it stands at present (December 1993). The following factors are being considered for inrolment in future versions of the methane sub-routine.

- Decay potential - as mentioned the model should have differentiated productivity with depth.
- Temperature dependencies of methanogens and methanotrophs are considered equal. Experiments should be carried out with literature values on Q_{10} values from similar soil environments.
- Physical limitations on gas transport in the soil are not considered in the model at present. Ideally, the model should incorporate physical factors that introduce contraints on the availability of e.g. O_2 and CH_4 at depth.
- The model should incorporate methane flux dependency on changes in nutrient and vascular plant status (once appropriate quantitative relationships become available).
- The effects of a possible changing plant cover should also be incorporated.

In conclusion, despite these uncertainties, the performance of the model seems with simple inputs to reproduce the most important interactions controlling methane flux in tundra environments. However, care should be taken in extrapolating the 2 x CO_2 result to produce changes in global tundra methane emissions. The model is applicable to wet/moist tundra areas only and, as discussed in Chapter 5 and above, even in these environments field (Svensson, 1983; Sebacher *et al.*, 1986; Crill *et al.*, 1988; Moore *et al.*, 1990; Bartlett *et al.*, 1992; Morrissey and Livingston, 1992; Whalen and Reeburgh, 1992; Christensen, 1993; Roulet *et al.*, 1993) and modelling studies (Roulet *et al.*, 1992a; Harriss *et al.*, 1993) suggest that methane fluxes are highly sensitive to soil moisture heterogeneity. The possible feedback effect on climate warming from possible increased tundra methane emission are therefore highly dependent on regional differences in soil moisture change. It should also be kept in

mind that the model results presented here is applicable only to somewhere between 1 and 2 x 10^{12} m^2 of wet and partly tussock/sedge tundra which means up to about one third of the global tundra areas (Chapter 2 and 3).

Disgarding the uncertainties, this work does indicate a potentially positive feedback on the anthropogenic greenhouse warming associated with increased methane emissions from tundra. Depending on the areal distribution of change in global tundra soil moisture regime this feedback could vary from < 1% increase of present global emissions to a significant 5% increase. The limited number of methane emission models available agree reasonably on the response. The bulk of uncertainty in assessing the possible feedback effect on greenhouse warming lies therefore with GCM soil moisture change predictions.

6.7. Summary

Tundra regions are predicted to experience a mean warming of approximately 4°C in winter and 2°C in summer. How this, combined with precipitation changes, will translate into soil moisture change will be of great importance to trace gas balance questions and whether methane emission might provide a significant feedback upon greenhouse warming. Very few model attempts to assess the latter question are presently available. This chapter explored one such attempt.

We extended the Meteorological Office single column model with soil thermodynamics and hydrology that allow it to incorporate the special behaviours of permafrost soils. A simple methane model integrating the effects on net methane emission of soil temperature, soil moisture and thaw depth was attached to the model. Initially the model was driven in a stand-alone mode forced by real weather data from the area in Alaska where measurements of methane emission was carried out (Chapter 4 and 5). The model showed itself capable of reproducing the measured soil environment satisfactorily. Temporal variations in methane flux over a seven-week period in the summer of 1991 were also reproduced by the model, providing confidence that,

however simple an approach taken, it seems to integrate the main controlling factors in a realistic way.

A number of sensitivity runs with the stand-alone mode of the model were carried out. An air warming of 4°C on average produced a 11% increase in thaw depth and a 13% increase in methane emission with no change in precipitation. With no change in the temperature regime, the methane emission was correlated linearly with precipitation. According to the stand-alone version of the model a mean 4°C increase and 10% more rain would cause a 21% increase in methane emission.

A full single column model run showed that the model was stable when being driven with true climatology and the full number of atmospheric layers. Here the model also reproduced general soil conditions on the tundra satisfactorily. A simple 2 x CO_2 scenario carried out with this model showed a dramatic response. The mean annual thaw depth increased with 42% and the soil moisture with 17%. The combination of slightly warmer and more moist soils and a larger soil volume produced a methane emission increase of 56%. Only one scenario was produced with the single column model and global significance of the result will depend on regional differences in soil moisture change. However, the fact that this model run incorporates the effect of warming on the seasonal cycle (i.e. a longer thaw season), as well as on the summer activity, makes the prediction without considering extrapolation more reliable than those produced using only one season's data.

The predictions made by various modes of the model vary from producing an insignificant < 1% increase in global methane emissions to a potentially important approximately 5% increase. The methane sensitivity studies available agree reasonably well on net emission response to a changing soil environment. The majority of the uncertainty rests with the GCM predictions of soil moisture change following global warming.

Chapter 7

A multigas analysis

7.1. Introduction

This chapter presents a limited dataset serving as basis for a short review of the tundra role in the atmospheric budgets of three other trace gases significant in climatic change. The data are from a simoultaneous analysis of CH_4, CO_2, N_2O and CO emission which I carried out on 21 August 1992 at the sites described in section 4.2 on the North Slope of Alaska. The methods of sampling and analyses are described in detail in Appendix 1. Samples were taken from aluminium dark chambers (Figure 7.1), a method different from the methane flux measurements described in Chapter 4 and 5, where transparent plexiglas chambers were used.

Figure 7.1. Chamber used for multigas sampling in place at a moss site (M1).

Table 7.1. Trace gas flux and environmental variables at nine stations in five different floristic units of Arctic tundra as measured on 21 August 1992. Note the different unit for CO_2 flux. The data in brackets are the molar equivalent of carbon. ND : not detectable.

	Soil temp. (°C)	Water table (cm)	Thaw depth (cm)	Organic mat. depth (cm)	CO_2 g/m²/day	CH_4 mg/m²/day	N_2O mg/m²/day	CO mg/m²/day
C1	9.3	-13	-48	>40	4.3 (1.2)	13.3 (10.0)	ND	0.7 (0.3)
C2	9.3	-13	-48	>40	4.0 (1.1)	81.1 (60.8)	ND	1.3 (0.6)
D1	3.6	10	-44	>30	3.1 (0.8)	104.3 (78.2)	ND	0.3 (0.1)
D2	3.6	11	-44	>30	0.6 (0.2)	24.8 (18.6)	ND	0.3 (0.1)
E1	8.9	-8	-65	>40	4.6 (1.2)	25.8 (19.4)	ND	0.9 (0.4)
M1	3.4	-4	-64	10	3.1 (0.8)	11.6 (8.7)	ND	0.5 (0.2)
M2	3.4	-7	-45	10	3.6 (1.0)	17.2 (12.9)	ND	1.0 (0.4)
T1	8.3	-8	-61	>40	16.0 (4.3)	181.5 (136.1)	ND	3.1 (1.3)
T3	8.3	-14	-50	>40	12.9 (3.5)	67.7 (50.8)	ND	1.9 (0.8)

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Table 7.2. Geographical extrapolation of ecosystem fluxes. Note the different unit for CO_2 flux.

	Moist tussock tundra 6.46 x 10^{12} m^2		**Wet meadow tundra** 0.88 x 10^{12} m^2		**Total tundra** 7.34 x 10^{12} m^2	
	Range	Mean	Range	Mean	Range	Mean
g $CO_2/m^2/day$	4.8-8.8	6.8	2.7-3.4	3.1	4.6-8.2	6.4
g CO_2-$C/m^2/day$	1.3-2.4	1.8	0.7-0.9	0.8	1.2-2.2	1.7
mg $CH_4/m^2/day$	36.9-69.1	53.0	46.1-53.2	49.6	38.0-67.1	52.6
mg CH_4-$C/m^2/day$	27.7-51.8	39.7	34.6-39.9	37.2	28.5-50.6	39.4
CH_4-C/CO_2-C		*0.022*		*0.045*		*0.023*
mg $CO/m^2/day$	0.88-1.61	1.24	0.58-0.71	0.64	0.8-1.5	1.2
mg CO-$C/m^2/day$	0.38-0.69	0.53	0.25-0.31	0.28	0.36-0.64	0.50

7.2. Carbon dioxide

The CO_2 release reported in Table 7.1 is a result of total respiration, which is a function of several respiratory processes such as microbial heterotrophic respiration, root respiration, dark respiration by green plants and soil fauna activity. It is analytically difficult to separate these processes, but various studies indicate that between one-third and two-thirds of total respiration originates from living roots (Heal *et al.*, 1981). By contrast, Svensson (1980) estimated this proportion to be lower (approx. 10%), finding instead that about 70% of the CO_2 released from a sub-arctic mire originates from microbial activity. Disregarding uncertainty about individual contributions, the total respiration can be compared with the amount of carbon fixed by primary production to give a measure of whether the system is a source or sink of atmospheric CO_2. This will also give some indication of the extent by which the system is either accumulating or losing dead carbon.

Table 7.1 illustrates that total CO_2 production is highest in the relatively warm aerated soils of the T, C and E units. C normally represents a wetter environment, but at the end of the thaw season the water table tends to drop (see Figure 4.4.a). The lowest emissions were found in the wetter and colder D and M units (Table 7.1). As described in section 4.2.1 D, M and C can be characterised as representative of the main botanical composition of wet meadow tundra and T, M and E of moist tussock tundra. Using these two subgroups of Arctic tundra the range of total respiration measured in the wet meadow tundra amounted to 0.16-1.17 g CO_2-C/m^2/day and 0.84-4.32 g CO_2-C/m^2/day for the moist tussock tundra. Data for the areal coverage of each unit within the two tundra subgroups can be used to obtain a figure for ecosystem flux. The framework used as basis for Table 7.2 was the same as used in section 4.4.2 for extrapolating tundra CH_4 flux. Given the limitation of the present data set, spatial and temporal extrapolation will not be attempted. However, a figure for landscape flux at the time and place of sampling can be calculated and compared with similar studies. Intertussock depressions (BH sites) were not sampled in these analyses due to the different chamber size which did not match the aluminium chambers used. The BH sites without vegetation are waterlogged and were

shown in Chapter 4 to have very low levels of microbial activity. In the following calculation they are assumed to have zero CO_2 emission.

When the mean fluxes are multiplied by the fraction of area covered by the floristic units, the total respiration of tussock tundra was calculated at 1.3-2.4 g CO_2-C/m^2/day with an average of 1.8 g CO_2-C/m^2/day (Table 7.2). The corresponding figure for wet meadow tundra is 0.7-0.9 g CO_2-C/m^2/day with average 0.8 g CO_2-C/m^2/day. In terms of wet meadow tundra this result is comparable to those obtained by Silvola and Heikkinen (1979) and Svensson (1980) (both approx. 0.6 g CO_2-C/m^2/day), respectively in a Finnish bog and Swedish sub-arctic mire. It is consistent also with the figure of 0.73 g CO_2-C/m^2/day for nocturnal tundra ecosystem respiration obtained with micrometeorological methods in the ABLE 3A expedition in Sub-Arctic Alaska (Fan *et al.*, 1992). However, it is significantly lower than the rate of approx. 1.9 g CO_2-C/m^2/day found at the IBP site on the coastal tundra near Barrow, Alaska (Heal *et al.*, 1981). The total respiration in the moist tussock tundra is significantly higher than the wet meadow tundra and more in the range of that reported by Heal *et al.*. Emission from the T sites on the moist tussock tundra even exceed the amount of CO_2 evolution reported for temperate forest soils (approx. 2 g CO_2-C/m^2/day) (Bouwman, 1990). The general pattern that dryer environments have higher rates of CO_2 evolution is not surprising, since aeration is a well-documented limiting factor for respiratory processes in the soil.

Total tundra flux in Table 7.2 was calculated with respect to the individual sizes of moist tussock tundra, 6.46 x 10^{12} m^2, and wet meadow tundra, 0.884 x 10^{12} m^2 quoted by Mathews and Fung (1987, Chapter 2). Given the larger coverage of moist tussock tundra, the total tundra mean estimate (1.7 g CO_2-C/m^2/day) comes out closer to that (1.8 g CO_2-C/m^2/day) rather than to the wet meadow tundra estimate (0.8 g CO_2-C/m^2/day).

Typical literature values for net primary production of the mentioned tundra ecosystem subgroups are 1.6 g C/m^2/day for moist tussock tundra and 0.7 g C/m^2/day for wet meadow tundra (Chapter 3, Shaver and Chapin, 1991). Comparing those figures with those for total respiration calculated above, a surplus of carbon loss over gain for both tundra types seems to be indicated. However, the significance of this result is

questionable given the relatively small amount of data and the unknown proportion of total respiration carried out by living plants, including roots. The latter will only have to be about 20% for the ecosystems to be in balance in terms of storage of dead organic carbon based on the results presented here. This notwithstanding, the result is in line with the findings by other studies discussed in section 3.1 which show the tussock tundra loosing carbon at a rate of 50-280 g C/m^2/yr (Grulke *et al.*, 1990; Oechel *et al.*, 1993). Even so, the possible rate of loss based on the data presented above would be significantly lower (about 20 g C/m^2/yr with an active season of 100 days).

7.3. Methane

The CH_4 fluxes measured in the multigas analysis were comparable to results reported in Chapter 4. The only exception to this is M2 where the measured flux of 17.2 mg CH_4/m^2/day was much higher than was obtained in monitoring work at the same sites (4±3 mg CH_4/m^2/day, Table 7.3). The flux was highest (8 mg CH_4/m^2/day) at M2 on the day of the monitoring closest to (and just before) the day where multigas sampling was carried out. This could be explained as the start of an episodic event of high flux at M2 in the days of multigas sampling. Such episodic events were discussed in section 4.2.1.

The fact that the fluxes generally are very similar to those measured with transparent plexiglas chambers in the monitoring work indicates that CH_4 flux measurements of the kind described here are insensitive to possible influences by light (see section 5.5.7).

The amount of organic carbon decomposing to CH_4 relative to CO_2 (the CH_4-C/CO_2-C ratio) was computed in Table 7.2 using the figures for ecosystem flux. The ratio is about 0.022 for tussock tundra and 0.045 for wet meadow tundra. The difference between these figures is consistent with the theory that wet anaerobic soil environments supports methanogenesis and has the effect of inhibiting both CH_4 oxidation and respiratory CO_2 producing processes. The ratios are in the range found in wet peat soils by Moore and Knowles (1989) in a study manipulating water table in laboratory columns. The total tundra calculation shows that about 2.2% of carbon flux to the

atmosphere is in the form of CH_4. This is a little higher than, but not conflicting with, the corresponding figure of 1.8% found by Svensson (1980).

Table 7.3. Methane emission as measured in this study compared to other studies at the same sites. Figures in mg $CH_4/m^2/day$.

	This study	Monitoring[a]	Other
C1	13.3	27±14	9.4[b]
C2	81.1	783±895	76.0[c]
D1	104.3	139±32	102.7[b]
D2	24.8	64±30	26.8[b]
E1	25.8	73±20	34.9[b]
M1	11.6	16±4	-
M2	17.2	4±3	8.5[b]
T1	181.5	141±65	-
T3	67.7	43±34	-

[a] Mean emission (± standard deviation) as measured in monitoring work during earlier part of August 1992 (Chapter 3). The high mean flux at C2 were due to an episodic event of exceptional high flux (up to 2.2 g $CH_4/m^2/day$).

[b] From the same dataset as [a] but the emission as measured on the day closest to that of the multigas sampling (Aug. 19 and Aug. 21 respectively).

[c] Flux measured on Aug. 20 in connection with sampling for isotopical analysis of CH_4 and CO_2 (manuscipt in preparation).

7.4. Nitrous oxide

Table 7.1 shows that no N_2O flux were obtained in this study. There were small changes in the chamber concentrations of N_2O but none rose above the uncertainty level of the flux calculation.

Mineralisation of organic N to N_2 and N_2O occurs through nitrification (oxidation of ammonium) and denitrification (reduction of nitrate) in the soil the latter being an anaerobic process. This suggests that wetlands could be a major source of N_2O to the atmosphere as they are in the case of CH_4. Indeed, relatively high rates of N_2O emission have been found in a variety of wetlands (Bowden, 1986). This has led to suggestions

that northern peatlands and tundra might account for some of the N_2O presently lacking in the estimated atmospheric N_2O budget (Schlesinger, 1991).

However, there is strong evidence contradicting the possibility that the tundra could be a major source of N_2O, in accordance with the results presented here: generally N mineralisation in Arctic soils is very low compared to soils of lower latitudes (Nadelhoffer *et al.*, 1992). The availability of nutrients is extremely low in the Arctic tundra and in order to overcome this, the ecosystems have adapted to a very slow flow of nutrients. This is the case for nitrogen in particular. An important buffer for nutrients between the living biomass and the atmosphere is the soil organic matter (SOM) where large amounts of nitrogen tend to accumulate. This further slows down the flow (Dowding *et al.*, 1981). In addition, the net N mineralisation that does occur has been shown to be at the lowest level at the height of the growing season (Giblin *et al.*, 1993). It may be a prolonged effect of this peak season shortage of mineralised N which is observed in this study.

The effect of the low mineralisation rates is reflected in the importance to tundra ecosystems of processes such as the direct uptake of amino acids from soils, N fixation and particularly N retranslocation in plants (Nadelhoffer *et al.*, 1992). Berendse and Jonasson (1992) shows how plant species in the tundra retranslocate up to 90% of their peak leaf nutrient content to storage organs before leaf senescence.

Together, the above forms a picture of ecosystems which at present are not likely to be losing significant amounts of nitrogen as N_2 or N_2O to the atmosphere. Furthermore the ratio of N_2O to N_2 might be very low. There is evidence that continuous wet soils have a low N_2O/N_2 ratio, while alternating flooding and draining greatly increases the N_2O flux (Bouwman, 1990). This is probably due to constantly wet soils allowing reduction to proceed fully to N_2. It is possible that the present moisture changes in the tundra are at such a small scale that the reduction to N_2 is the prevailing process. Possibly more dramatic seasonal variations following climatic change could increase the N_2O/N_2 ratio as well as increase the total net N mineralisation, thereby causing a feedback effect

on global warming. However, much more data is needed in order to assess this question with greater certainty.

Martikainen *et al.* (1993) investigated the possible response of N_2O emission from Finnish peatlands with different nutrient status to a warmer and drier climate. Only the nutrient-rich peatlands showed an increasing emission following drying but the scale was so small that they concluded northern peatlands are unlikely to exert a significant climate feedback from N_2O emissions. A recent modelling study investigating the global soil production of N_2O also found the lowest potential emission at high northern latitudes (Bouwman *et al.*, 1993).

7.5. Carbon monoxide

As shown in Table 7.1, the levels of CO emission measured at all sites ranged between 0.3 and 3.1 $mg/m^2/day$. The higher levels were recorded at tussocks, the lowest in depressions.

Broadly speaking, emission of CO from natural soil/vegetation environments can have three sources: direct release by plants, oxidation of hydrocarbons (isoprenes and terpenes) released by plants, and chemical production in soils (Logan *et al.*, 1981; Bouwman, 1990).

The process behind direct release of CO by plants is not well understood, but Seiler and Giehl (1977) showed how the release is highly light-sensitive. They found no CO emission during dark incubations of plants *Vicia faba* and *Platanus acerifolia*. The oxidation of isoprene released from plants has also been shown to be light-sensitive (Zimmerman *et al.*, 1978). Investigating the relevant atmospheric reactions, Zimmerman (1978) concluded that the appearance of CO from oxidation of plant hydrocarbons would be delayed, relocated, or most likely prevented, if concentrations of NO (a necessary precursor for the reactions) were lower than 50 ppt. Such high concentrations would appear only in areas of high pollution or intensity of lightning (mean Alaskan NO and NO_x values are about 8.5 and 25 ppt respectively (Sandholm *et al.*, 1992)). Since neither of these situations are attributable to the tundra surveyed here, and the measurement

technique involved dark chambers, both direct release by plants and oxidation of plant hydrocarbons are assumed to be insignificant processes in terms of the CO production measured.

This leaves CO emission by soil. Production of CO in soil has been shown to be a chemical process as opposed to soil CO consumption which is microbially mediated (Conrad and Seiler, 1985). No studies of CO flux in northern wetland soils are available for comparison, but Conrad and Seiler (1985) found CO emissions from sub-tropical and temperate soils to be in the range of 0.2-1.1 mg/m^2/day and 0.08-1.2 mg/m^2/day respectively. These are overlapping but generally lower ranges than the results of this study show (see below). The same authors found CO production to be decreasing with falling soil temperatures, increasing soil moisture and falling organic carbon content in the soil. In an earlier study of arid soils they showed how CO emission turned into deposition when the soils were irrigated (Conrad and Seiler, 1982). At first inspection these results does not seem in line with the relatively high emissions found in the cold wet and moist tundra soils. However, if the high organic carbon content of tundra soils is considered, the results seem more compatible. Conrad and Seiler (1985) proposed a relationship where CO production is a function of soil surface temperature and soil organic carbon content. With soil surface temperatures for tundra soils of typically around 10°C, the soil organic carbon content would have to be approximately 10% for the relationship to produce the overall estimated mean flux of 1.2 mg/m^2/day observed in this study (Table 7.2). The organic carbon content is generally higher than 10% in peaty soils of Arctic tundra. This might compensate for the soils being wetter than those investigated by Conrad and Seiler. Hence, by incorporating a moisture variable (and testing it in drained and undrained northern peat soils) the relationship developed by Conrad and Seiler could also prove valid in wet northern soils.

The CO emission in this study is linearly correlated with the CO_2 emission (r^2 = 0.89, n = 9). Since most tropospheric reactions involving CO and CO_2 are oxidising processes leading to CO loss and CO_2 gain (Crutzen and Zimmerman, 1991), and since the relative concentration of CO is far too low to influence the concentration of CO_2, the

relationship seems unlikely to be due to chemical reactions between the two species in chambers or sampling vials. In addition, the measurements of ambient CO (mean: 108 ppb) did not differ from values cited in the existing literature for the Alaskan Arctic (90-110 ppb) (Khalil and Rasmussen, 1990b; Harriss *et al.*, 1992). The only possible explanation for the correlation is that the two processes have very similar environmental constraints. In other words, relative drier, warmer, and more organic soils provide the best conditions for CO (and CO_2) production. This pattern also corresponds well with the findings of Conrad and Seiler (1985) mentioned above.

A crude extrapolation of CO flux to global tundra based on the overall mean emission calculated in Table 7.2 provide figures of 0.6-1.1 Tg/yr (based on a 100 day active season). This presents a modest, but so far unaccounted, addition to the global soil CO emission of 3-30 Tg/yr calculated by Conrad and Seiler (1985) and quoted in recent reviews (Bouwman, 1990).

7.6. Summary

Fluxes of trace gases CO_2, CH_4, N_2O and CO were measured on 21st August 1992 at nine sites representative of Arctic tundra on the North Slope of Alaska.

Total respiration by wet meadow tundra was calculated at 0.7-0.9 g CO_2-C/m^2/day and for moist tussock tundra at 1.3-2.4 g CO_2-C/m^2/day. These figures are in accord with most other measurements in similar environments. A comparison of the results with literature data on NPP in comparable ecosystems seems to indicate a small loss of carbon from the tussock tundra surveyed. Although this is a highly uncertain conclusion given the small data set, it supports other recent findings from the same study area.

The CH_4 fluxes measured using dark chambers compares generally well with the results presented in Chapter 4 (using transparent chambers). This indicates that closed chamber flux measurements are insensitive to the influence of light in accord with the conclusion drawn in section 5.5.7. The amount of carbon decomposed to CH_4 relative to CO_2 in wet meadow tundra was about twice that of the moist tussock tundra. The tundra showed overall 2.2% of carbon flux to the atmosphere to be in the form of CH_4.

Zero N_2O flux was measured in this study. This seems a reasonable result given the general very low N mineralisation rates in Arctic tundra ecosystems. Furthermore, the measurements were carried out in the high season, which has been shown in literature to be the time of lowest mineralisation.

A significant CO flux was measured. The emission rates were in the upper range of that reported for tropical and temperate soils. Although the literature suggests that the wet conditions of tundra soils should not be favourable for CO production, the emission seemed to be environmentally constrained in a way comparable to soils of lower latitudes. Emission of CO and CO_2 were linearly correlated, which suggests very similar controls on the production (drier, warmer and more organic soil environments increases emission).

Chapter 8

Conclusions

In Chapter 1 I identified a number of questions. The thesis has been focused around exploring those based on both my own work and information from the literature. The following is a summary of answers to the questions found during the course of the thesis work.

What are the general physical and biological characteristics of tundra and how are those interlinked with climate?

This is obviously a very broad question to which a number of answers of no relevance to this thesis could be given. Of importance in the present context are the primary control on ecosystem functioning in the Arctic: net radiation. Net radiation, or energy input, in combination with the albedo effect, are main determinants of global tundra distribution. Good approximations of the areal limits of major vegetational groups (e.g. the tree line) can be made from "isolines" of a certain number of degree days above 0°C or the mean July temperature isotherm.

The low energy input cause annual mean temperatures to be below 0°C. This also forms basis for the existence of permafrost, a zone of permanently frozen ground below the soil surface. Tundra regions receive very little precipitation but the low energy input, causing low rates of evaporation, combined with the presence of permafrost, which inhibits drainage, result in generally wet soil conditions in the tundra. The wet soil conditions promotes the accumulation of peat in the ground (see below). Peat, snow, vegetation and water bodies all have important effects on the depth of the seasonally thawed soil layer above the permafrost, the active layer. This layer is where biogeochemical cycling, subject of study in this thesis, takes place. Although the tundra biome generally is young, soils and vegetation typically not dating back more than 12,000 years, its large areal coverage (between 5.7 and 7.4 x 10^{12} m^2) and tendency to

accumulate carbon makes it a potential important player in global carbon cycling thus also providing possible feedback effects on global climate change.

How does these characteristics affect carbon cycling in tundra ecosystems?

The low energy input cause net primary production to be low in the tundra compared with most other biomes. A mean figure for tundra is 65 g $C/m^2/yr$ but the range varies between 25 g $C/m^2/yr$ for dry heath tundra to a maximum of 1000 g $C/m^2/yr$ in some shrub tundra environments. However low the mean figures for net primary production are, tundra soils have accumulated considerable amounts of carbon during the Holocene. This is due to the wet and cold conditions exerting stronger limitations on total decomposition relative to production. There are controversy as to the size of the tundra soil carbon pool but areally weighted mean figures for total tundra varies between 10 and 20 kg C/m^2. The tundra as a whole probably still accumulates carbon but it has recently been suggested that particular regions of tussock tundra has changed to become atmospheric carbon sources in response to recent climatic change.

Disregarding the uncertainty about present tundra carbon balance, the tundra soils remain relatively wet and anaerobic decomposition in waterlogged parts of the soils plays an important role in tundra carbon cycling. Methane is the reduced C-compound resulting from anaerobic decomposition and tundra regions are generally known as substantial sources of atmospheric methane. Not all of the methane produced at depth in tundra soils reaches the atmosphere. If an aerobic zone above the area of methane production is present, then methane oxidation may be in active controlling net flux to the atmosphere. Net methane emission (and total carbon loss) to the atmosphere is therefore highly dependent upon the climatically controlled water balance of tundra soils as well as on temperature and other factors discussed in this thesis. The main focus here is the control on methane emission which has given rise to the next couple of questions.

What are the present rates and isotopical signature of methane emitted from true arctic tundra?

I measured methane emission at 22 stations representative of tussock and wet meadow tundra on the North Slope of Alaska in the summers of 1991 and 1992. Six floristic units showed significant different seasonal mean fluxes varying from 112±72 mg CH_4/m^2/day for *Carex* sites to 0.6±0.4 mg CH_4/m^2/day for inter-tussock depressions. Episodic events of very high emissions (up to 2.2 g CH_4/m^2/day) were repeatedly measured and might account for about 25% of global tundra flux.

Using information on the areal coverage of the floristic units and mean flux figures from 1991 and 1992 tussock tundra emission on the North Slope of Alaska were calculated at 15-29 mg CH_4/m^2/day. Similar figures for wet meadow tundra were 88-100 mg CH_4/m^2/day. This yields total Alaskan flux figures of tussock tundra: 0.2-0.4 Tg/yr and wet meadow tundra: 0.9-1.1 Tg/yr. Assuming Alaskan tussock and wet meadow tundras are representative for these subcategories a total global tussock tundra flux of 1.5-2.6 Tg/yr and wet meadow flux 7.8-9.0 Tg/yr are calculated.

In order to assess global tundra emission, assumptions will have to be made about emission from other tundra types such as shrub-, sedge-dwarf shrub tundras and also polar semidesserts which are included in global tundra emission estimates. Flux data from these environments are not readily available. Instead, for comparative reasons, a widely used crude scheme for extrapolating global tundra flux was employed. This yielded a global tundra methane emission of 18-30 Tg/yr based on my own data. A corresponding estimate derived on the basis of all available flux data in the litterature arrived at 19.5±5 Tg/yr. Combining these two estimates with fluxes from remaining parts of northern wetlands and tundra lakes north of 50°N derives a total estimated flux of 19-35 Tg/yr.

A limited data set on the isotopical signature of methane emitted from the true tundra sites on the North Slope were obtained in 1992. Mean $\delta^{13}CH_4$ was -63.9±3.9‰. Radiocarbon in emitted CH_4 varied between 105 and 114 pMC and 113 to 115 for CO_2.

How does this emission compare with more extensively surveyed tundra-like environments in the Sub-Arctic?

Tundra methane emission is highly variable both between sites and temporally at the same sites. It is therefore difficult to compare immediate rates of emission from two different regions. However, seasonal ranges and mean emission from equivalent tundra subcategories can be compared. Wet meadow tundra as measured in this study falls within the range found by Whalen and Reeburgh in four-year time series of measurements in sub-arctic interior Alaska. It is also in line with other studies of wet tundra in sub-arctic Alaska.

Substantial differences are found, however, in terms of tussock tundra emission. Studies in sub-arctic tundra environments as well as in the same area as the present found lower tussock fluxes. This is probably because tussock tundra environments covers a wide range of moisture conditions. A commonly used scheme for extrapolating methane flux only operates with two tundra subcategories; wet and dry/moist tundra. The latter grouping covers very large differences in soil, vegetation and moisture characteristics. For example, tussock tundra, often used in extrapolating dry/moist tundra flux, is widespread under very different environmental conditions which probably gives rise to different methane emission figures obtained in this environment.

Very few data are available on isotopical signatures of methane emitted from true tundra environments. In a previous study of sub-arctic tundra methane flux $\delta^{13}C$ values of -65.8±2.2‰ was found which is in agreement with the results of -63.9±3.9‰ presented here.

Do the characteristics of methane emitted from true arctic tundra correspond with the assumptions made about it in recent attempts to estimate the atmospheric methane budget?

Methane emission from northern environments was until recently believed to be significantly higher than what this study has shown. Estimates for total northern wetland including tundra flux was estimated up to 100 Tg/yr as recent as in the late 1980s. Following, a rapidly increasing number of flux studies this figure is now estimated at less

than 50 Tg/yr which is in agreement with the extrapolations carried out in this thesis. The most extensive and up-to-date attempt to estimate the global methane budget operated with a northern wetland and tundra emission of 35 Tg which is in the upper range but consistent with my findings. In the same study isotopical signatures of varies sources and fractionation rates were used to constrain the budget. In the preferred scenario methane from tundra environments were assumed to have an $\delta^{13}C$ value of -61‰ which is in agreement with the results pressented here.

In short, this study supports atmospheric methane budgets which also operate with a global tundra and northern wetland methane emission figure of somewhere between 15 and 40 Tg/yr and $\delta^{13}C$ values of this methane between -60 and -67‰.

What are the controlling factors on net methane flux to the atmosphere?

Factors having significant spatial or temporal controls on net methane emission from tundra soils as identified in this study include:

- soil temperature (and thaw depth)
- soil moisture
- soil pH
- nutrient availability
- organic material
- vascular plant cover

Together they form a complex pattern of controls. However, while it is acknowledged that other controlling factors (such as vascular plants) might have some influence it is postulated that controls on seasonal variations in net flux may be considered a function of only three factors: soil temperature, soil moisture and thaw depth.

Can the most important of these factors be given priorities and quantified in a way that allows predictive models to reproduce seasonal variations in methane flux?

In this study a model was developed that builds on the assumption made in the preceding answer, that soil temperature, soil moisture and thaw depth are the most important factors influencing temporal variations in net methane flux. A simple integration of these factors in a methane routine combined with a physical model developed that incorporates the behaviour of permafrost soils was used to test the question.

The model showed capable of reproducing seasonal variations in net flux on the North Slope of Alaska. The validation run was limited timewise but it does seem possible to reproduce seasonal variations in flux with such a simple model.

How will the modelled tundra methane emission respond to climate change scenarios predicted by GCMs and what are the major uncertainties in such predictions?

The model was tested for its response to climate change scenarios in different modes. A number of sensitivity runs were carried out. An air warming of 4 °C on average produced a 13% increase in thaw depth and a 13% increase in methane emission with no change in precipitation. With no change in temperature, the methane emission had a linear response to changes in precipitation. According to this version of the model a mean 4°C increase and 10% higher precipitation in summer would cause a 21% increase in methane emission.

A full multi-year single column run with the model showed a more dramatic response to a simple 2 x CO_2 scenario. The mean annual thaw depth increased with 42% and the soil moisture with 17%. The combination of slightly warmer and more moist soils and a larger soil volume produced methane emission increase of 56%.

It is important to note that the uncertainty in these predictions are highly dependent upon the representation of soil environment change (in particular moisture) in the GCMs. Most methane models agrees relatively well on the general mechanisms behind methane flux and its possible response to changing soil temperature and moisture. The bulk of the uncertainty in any prediction of methane emission response to climatic change lies in the

predictions of soil environmental change. Also, it is probably unlikely that all tundra regions will experience the same changes in climate. This could be an important factor limiting the potential feedback effect on climate change.

What are the general roles of tundra regions in the atmospheric budgets of other trace gases (CO_2, N_2O, CO) with relevance to climatic change ?

Tundra ecosystems are generally known as overall sinks for atmospheric carbon dioxide. However, it has recently been claimed that tussock tundra on the North Slope of Alaska has changed to become an atmospheric carbon dioxide source. The limited data on carbon dioxide flux in my study does not contribute much to this discussion. The data I obtained indicates, when balanced against primary production data from the same region, that the system is roughly in balance with a slight tendency to a net loss at the dryer sites.

I measured no N_2O flux at my tundra sites on the North Slope. This is not surprising since tundra ecosystems in general are known to be highly nutrient limited and there are therefore little nitrogen in surplus for potential mineralisation in the form of N_2O.

Contrary to N_2O I measured a significant CO emission in the order of 0.3-3.1 mg $CO/m^2/day$. This is apparently the first CO flux measurements from tundra environments. Most soil CO emission reported in the literature is associated with arid soils and the flux found here may therefore seem surprising. However, a high organic carbon content have also been shown to increase CO emission which could form part of the reason why tundra soils show a significant flux.

Are there reasons to believe that the dynamics of these gases in tundra environments could provide feedback effects upon climatic change?

A possible feedback from increased emissions of carbon dioxide from tundra environments is much discussed. It is closely linked to the modelling question about possible increased methane emissions. Undoubtedly the tundra will lose carbon as a result of climate warming. The possible fertilisation effect of increased ambient CO_2

concentrations and higher temperatures have not been shown capable of counterbalancing the loss associated with increased decomposition following warming. The question is then whether the increased release of carbon will be in the form of methane or carbon dioxide or possibly both. Our model predicts increased methane emission following warming even if this is associated with a slight drying of the soils. If the soil environment warms with a slight decrease or no change in moisture the indications are that both gases might increase in emission. If it becomes substantially dryer then a significant increase in carbon dioxide and decrease of methane emission would be expected. The net result in terms of radiative forcing of climate is very difficult to assess but the two processes could end up outweighing each other.

The possible feedback effect of tundra nitrous oxide emission on climate change is estimated as negligible. This is due to the general very low nutrient availability in most tundra ecosystems. However, the nutrient status of tundra soils also have consequences for other processes such as microbial methane oxidation. A changed nutrient cycling in tundra as a consequence of climate warming may therefore have important secondary effects, but very little is as yet known about that.

The CO emission measured in this study is the first reported from tundra soils and the environmental controls on CO flux is very poorly known. There seemed to be some resemblance between CO and CO_2 response to spatial differences in environmental conditions. However, given the small dataset I will not attempt to make suggestions as to how CO flux might change following changing climate.

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Appendix 1. Materials and methods

Methane flux measurements in Alaska (Chapter 4 and 5).

Net CH_4 fluxes were determined by a static chamber technique using aluminium bases and plexiglas covers sealed by water-filled channels. The chamber areas were 0.075 m^2 except at the BH sites where they were 0.023 m^2. Four duplicate 8 ml samples were taken at maximum 20 minute intervals with glass syringes, and methane was analysed at Toolik Lake Field Station using a gas chromatograph (Shimadzu GC-8A) equipped with a flame ionisation detector and a molecular sieve column. Ultrapure nitrogen was used as carrier gas. Methane flux was calculated from chamber size and the linear change of CH_4 concentration in the chamber with time. The minimum detectable flux varied between 0.2 and 1.2 mg CH_4/m^2/day depending on chamber volume:area ratios. Fluxes lower than the minimum detectable flux were considered zero. Soil temperatures at 1, 3, 5, 7, 9, 11 and 13 cm depth were obtained with a thermistor string and hand-held thermometer (Omega 866). Water table position was measured at each station in wells relative to soil surface and thaw depth was determined by inserting a steel rod to the freezing horizon. Soil pH were measured in slurries of soil and distilled water.

Soil core experiment (sections 4.4.4 and 5.4).

Cores were taken by cutting the peat in a circular motion with a diameter of 10.3 centimeter determined by core tubes which were gradually pushed down following the knife. Care was taken to avoid pressurising the peat. The tubes were 50 cm long and at a core depth of 45 cm the tubes were hermetically closed with a lid on the top. The pressure in the tubes made it possible to pull the soil out with limited cutting at the base of the soil cores. The soil were inundated when cores were taken and were kept so by transferring them to large buckets (45 cm height) filled with bog water. Since the tubes were open at the bottom the water table established itself at a natural level very close to the soil surface.

Net methane flux measurements were carried out by a closed chamber technique. During flux sampling a plastic lid equipped with a rupper septum was put on the cores. The flux was then calculated from the concentration change in the chamber with time and the ideal gas law. The enclosed air volume varied between 0.37 and 0.62 litres. Samples were normally taken at 0, 10 and 20 minutes after lid instalment. Samples were taken up to 45 minutes after lid instalment in cores with very low emission. Four-millilitre samples were taken with plastic syringe and transferred immediately to two-millilitre glas ampoules with silicone septa. Gas analyses were carried out on a Varian 3400 gas chromatograph equipped with a flame ionisation detector and autosampler, Varian 8100 (which injected 0.5 mililitre gas samples from the ampoules). Ultrapure nitrogen was used as carrier gas.

Holes in the core tubes at 10, 20 and 30 centimetres depth below soil surface equipped with rubber stoppers allowed for samples of pore water to be taken. A double-needle vacuum system (designed for human blood sampling) were used to suck between 2 and 5 mililitre porewater samples into 10-mililitre vacutainers. The exact water volume were determined by weighing the vacutainers before and after sampling. The vacutainer headspace was filled with nitrogen and shaken for two minutes to allow equilibrium to establish. Three-mililitre samples were then taken from the headspace and transferred to ampoules to be analysed as above. Porewater concentration of methane was calculated from the measured concentration, water volume and the Bunsen Ratio for methane (corrected for given temperature).

Multigas analysis (Chapter 7).

Measurements of trace gas flux were made on 21 August 1992 at the sites described in section 4.2. Net gas fluxes were determined by a technique developed by Whalen & Reeburgh (University of California, Irvine). Closed aluminium chambers with permanent aluminium bases sealed by water-filled channels were installed for a period of 45 minutes. The chamber areas were 0.075 m^2 and volumes varied between 21.9 and 31.3 litres. Eight 50 millilitre samples were taken at 0, 5, 10, 15, 20, 25, 30 and 45

minutes after chamber instalment with a glass/graphite syringe and transferred to evacuated 20 millilitre glass vials sealed with silicone stoppers.

Samples were brought to Fairbanks and analysed within two days of sampling. The analysis of CO_2 was carried out using a Shimadzu GC-8A GC equipped with a thermal conductivity detector. Shimadzu mini-2s fitted respectively, with a flame ionisation detector and an electron capture detector were used for analysis of CH_4 and N_2O. CO was analysed on a Trace Analytical model RG3A reduction gas analyser.

Flux was calculated on the basis of the concentration change of the gases in the chamber with time and the ideal gas law. Since dark chambers were used, the CO_2 flux is a measure of total respiration by plants and soil.

Soil temperatures adjacent to the flux chambers were determined at seven depths in the upper 13 centimetre of the soil by means of a portable thermistor string. The position of the water table was measured in wells, and thaw depth was determined by inserting a steel rod to the freezing horizon. A 40-50 centimetre hole was dug at each station to determine the depth of the organic layer.

Isotopical analyses (section 5.3).

Gas sampling for isotopical analyses was carried out with a modified version of the aluminium chambers used for multigas analysis. In order to carry out the analyses a large volume of gas had to be sampled. To avoid pressurising the sample-air a loose plastic bag was sealed to a hole in the chamber allowing air to penetrate into the chamber-space without contaminating the air being sampled. Sample air was transferred to sealed aluminium bags and shipped to Germany.

Hydrogen and stable carbon isotope analyses were carried out by Marcus Thom, Institute for Environmental Physics at University of Heidelberg. Note the deuterium values might be slightly too heavy, due to contamination by stray water, particularly when the samples were small. For details of methodology see Thom *et al.* (1993). Radiocarbon analyses on graphite targets, prepared in Heidelberg from the CH_4 and

CO_2 samples, were carried out through Accelerator Mass Spectrometry by Professor Bonani's group at Eidgenössische Technishe Hochschule (ETH) in Zürich.

Flux measurements at Abisko, Northern Sweden (section 4.3 and 5.5.2).

Methane flux at Stordalen were measured using a static chamber technique. The chambers were the same tubes as used in the soil core experiments cut up in 30 cm sections. Flux was calculated from the concentration change in chambers with time as described above. Samples were taken in plastic syringes and brought to the laboratory at Abisko Field Station and analysed within six hours of sampling.

Analysis of CH_4 was carried out on a Perkin-Elmer F11 gas chromatograph equipped with a FID detector and a Chromosorb 103 80/100 column. Ultrapure nitrogen was used as carrier gas.

Appendix 2. Alaskan methane flux data

The following tables are from the data files on methane flux obtained on the North Slope of Alaska in 1991 and 1992. See Chapter 4 for site explanations. The data from 1991 are under julian days (166 = 15 June, 242 = 29 August). The 1992 data are under dates in August 1992.

CH4 flux (mg/m2/day)

Julian day	C1	C2	C3	Mean C	Standard dev.	St. error
166	12.25	28.17	21.96	20.793333	8.023866483	4.63258
167	13.58	30.81	13.46	19.283333	9.98256647	5.76344
169	20.65	35.37	18.21	24.743333	9.28347636	5.35982
171	37.24	57.39	34.53	43.053333	12.48963704	7.2109
174	68.65	143.41	84.3	98.786667	39.4292028	22.7645
180	95.78	222.89	108.34	142.33667	70.04332968	40.4395
188	176.19	1528.02	85.22	596.47667	808.021424	466.511
189	116.09	712.26	63.04	297.13	360.4903096	208.129
190	73.1	305.14	61.16	146.46667	137.5447597	79.4115
192	43.01	480.44	44.57	189.34	252.1012017	145.551
194	42.53	1054.41	43.91	380.28333	583.8112265	337.064
195	29.88	195.88	44.33	90.03	91.95307227	53.0891
205	21.72	105.4	40.82	55.98	43.85150853	25.3177
212	19.67	151.58	46.68	72.643333	69.68237247	40.2311
214	17.56	104.8	41.05	54.47	45.14174011	26.0626
216	9.17	64.58	29.37	34.373333	28.04179084	16.1899
218	8.56	48.44	25.6	27.533333	20.01017075	11.5529
220	7.99	44.72	30.62	27.776667	18.52934519	10.6979
238	3.52	20.94	13.19	12.55	8.727617086	5.03889
240	4.49	19.07	11.94	11.833333	7.290585253	4.20922
242	4.66	21.26	14.22	13.38	8.331818529	4.81038
1992 Aug. 3	51.32	1587.55	40.47	559.78	890.0914618	513.895
5	32.92	2228.17	38.67	766.58667	1265.771561	730.794
7	28.28	206.29	31.56	88.71	101.8404728	58.7976
9	20.95	172.53	31.31	74.93	84.68265702	48.8916
11	19.71	327.7	25.71	124.37333	176.1116124	101.678
19	9.43	181.28	19.28	69.996667	96.4999525	55.7143
Observed mear	36.6259259	373.277778	39.3896296	149.76444		
Standard dev.	39.7617373	567.82865	23.7320224	199.15351		
Maximum	176.19	2228.17	108.34	766.58667		
Median	20.95	151.58	34.53	72.643333		

CH4 flux (mg/m2/day)						
Julian day	D1	D2	D3	Mean D	Standard dev.	St. error
166	49.88	3.44	0	17.773333	27.85833687	16.084
168	33.41	6.8	2.53	14.246667	16.73269952	9.66063
170	61.39	13.33	8.51	27.743333	29.23836065	16.8808
172	82.47	32.18	27.86	47.503333	30.35895969	17.5278
175	86.01	28.73	39.37	51.37	30.46718891	17.5902
180	85.71	38.46	45.25	56.473333	25.54629197	14.7492
187	147.24	47.92	61.65	85.603333	53.81855845	31.0722
189	142.84	35.88	76.43	85.05	53.99850646	31.1761
191	89.94	41.61	66.44	65.996667	24.16804985	13.9534
193	87.73	31.52	66.44	61.896667	28.37908443	16.3847
195	93.28	29.75	69.46	64.163333	32.09448914	18.5298
205	111.85	49.56	109.56	90.323333	35.32064599	20.3924
212	120.09	25.39	83.13	76.203333	47.7284667	27.556
213	162.43	28.75	80.17	90.45	67.4302929	38.9309
215	131.43	29	73.96	78.13	51.34216493	29.6424
216	121.61	16.09	57.95	65.216667	53.13399038	30.6769
217	127.33	18.96	53.04	66.443333	55.41435945	31.9935
219	160.64	17.03	44.95	74.206667	76.14408994	43.9618
238	133.38	19.69	35.01	62.693333	61.69383465	35.619
240	125.53	10.47	19.87	51.956667	63.88948688	36.8866
242	137.34	13.86	29.67	60.29	67.19386802	38.7944
1992 Aug. 1	205.51	116.8		161.155	62.72744256	44.355
3	122.61	69.68		96.145	37.42716193	26.465
5	139.68	80		109.84	42.2001327	29.84
7	138.12	71.08		104.6	47.40443861	33.52
9	137.33	45.16		91.245	65.17403202	46.085
11	127.92	43.07		85.495	59.99801038	42.425
19	102.72	26.82		64.77	53.66940469	37.95
Observed mear	116.622143	35.3939286	50.0595238	71.677976		
Standard dev.	36.5676312	24.8126366	28.8331973	29.264424		
Maximum	205.51	116.8	109.56	161.155		
Median	124.07	29.375	53.04	66.22		

CH4 flux (mg/m2/day)						
Julian day	E1	E2	E3	Mean E	Standard dev.	St. error
166	4.36	5.05		4.705	0.487903679	0.28169
168	3.87	7.12	3.12	4.7033333	2.126225137	1.22758
170	6.63	11.81	3.59	7.3433333	4.156168107	2.39956
172	13.59	26.6	7.47	15.886667	9.769607635	5.64049
175	17.46	31.24	6.56	18.42	12.36797477	7.14065
180	22.31	59.54		40.925	26.32558546	15.1991
187	31.56	42.47	15.7	29.91	13.46105865	7.77175
189	24.05	42.2	13.67	26.64	14.44026662	8.33709
191	16.54	33.16	8.43	19.376667	12.60667416	7.27847
193	21.66	37.46	7.45	22.19	15.01201852	8.66719
195	16.27	24.02	6.74	15.676667	8.65526622	4.99712
205	29.32	47.03	11.96	29.436667	17.53529108	10.124
212	27.26	54.52	13.14	31.64	21.03483777	12.1445
213	30.59	63.67	15.23	36.496667	24.75429121	14.2919
215	31.93	69.15	15.55	38.876667	27.46692799	15.858
216	25.41	52.01	14.14	30.52	19.4452642	11.2267
217	28.9	45.45	15.78	30.043333	14.86800704	8.58405
219	27.21	39.23	9.16	25.2	15.13543194	8.73845
238	28.04	23.05	7.88	19.656667	10.49963968	6.06197
240	16.8	18.3	12.87	15.99	2.804157628	1.61898
242	27.09	17.75	16.53	20.456667	5.776931135	3.33531
1992 Aug. 1	104.12	132.67		118.395	20.1878986	14.275
3	80.61	101.05		90.83	14.45326261	10.22
5	76.07	87.48		81.775	8.068088373	5.705
7	75.47	72.22		73.845	2.298097039	1.625
9	70.73	81.67		76.2	7.735748186	5.47
11	71.42	74.34		72.88	2.064751801	1.46
19	34.93	29.98		32.455	3.500178567	2.475
Observed mear	34.4357143	47.5085714	10.7878947	36.802619		
Standard dev.	25.9421015	29.9367772	4.33908923	28.530771		
Maximum	104.12	132.67	16.53	118.395		
Median	27.235	42.335	11.96	29.673333		

CH4 flux (mg/m2/day)						
Julian day	BH1	BH2	BH3	Mean BH	Standard dev.	St. error
166	0	0	0	0	0	0
168	0	0	0	0	0	0
170	0	0	0	0	0	0
172	0	0.39	0	0.13	0.225166605	0.13
175	0	0.86	0.22	0.36	0.446766158	0.25794
180	0.23	1.58	0	0.6033333	0.853600219	0.49283
187	1.25	1.88	0	1.0433333	0.956887315	0.55246
189	0.5	-0.28	0	0.0733333	0.395137107	0.22813
191	0.3	0	0	0.1	0.173205081	0.1
193	0.7	3.67	0	1.4566667	1.948495146	1.12496
195	0.17	1.76	0	0.6433333	0.970790056	0.56049
205	0.31	0.24	0.12	0.2233333	0.096090235	0.05548
212	0.14	2.94	0.14	1.0733333	1.616580754	0.93333
213	0.27	1.28	0.18	0.5766667	0.610764548	0.35263
215	0.27	0.2	0.83	0.4333333	0.3453018	0.19936
217	0.47	0.58	0	0.35	0.308058436	0.17786
219	0.1	1.13	0	0.41	0.625539767	0.36116
238	0.4	7.29	0.39	2.6933333	3.980833246	2.29834
240	0	3.49	0.55	1.3466667	1.876441668	1.08336
242	0.15	1.12	0	0.4233333	0.60797478	0.35101
1992 Aug. 1	1.37	7.46	0	2.9433333	3.971074582	2.2927
3	0.37	3.5	0	1.29	1.922836447	1.11015
5	2.25	11.45	0.25	4.65	5.973273809	3.44867
7	0.34	2.37	0.24	0.9833333	1.201929005	0.69393
9	1.17	4.82	0.22	2.07	2.428476889	1.40208
11	0.96	4.5	0.49	1.9833333	2.192129862	1.26563
19	0.44	2.59	0.74	1.2566667	1.164402565	0.67227
Observed mear	0.45037037	2.40074074	0.16185185	1.004321		
Standard dev.	0.53164101	2.7827379	0.24076859	1.0899972		
Maximum	2.25	11.45	0.83	4.65		
Median	0.3	1.58	0	0.6033333		

Q4 1

CH4flux (mg/m2/day)						
Julian day	M1	M2	M3	Mean M	Standard dev.	St. error
166	2.12	0	2.5	1.54	1.34714513	0.77777
168	3.64	0.46	4.71	2.9366667	2.210573078	1.27627
170	5.41	0	6.19	3.8666667	3.371265835	1.9464
172	6.3	-0.22	7.33	4.47	4.094178794	2.36378
175	12.59	0.69	16.19	9.8233333	8.111925378	4.68342
180	14.83	1.84	16.1	10.923333	7.891985386	4.55644
187	11.74	2.41	14.82	9.6566667	6.4619837	3.73083
189	10.09	0.68	14.45	8.4066667	7.037644019	4.06319
191	9.39	0.31	15.4	8.3666667	7.596869969	4.38605
193	8.18	0.98	12.66	7.2733333	5.892548967	3.40206
195	11.1	0	19.19	10.096667	9.634263508	5.56234
205	13.65	1.17	19.74	11.52	9.466461852	5.46546
212	17.74	2.85	23.97	14.853333	10.85187695	6.26533
213	17.76	2.36	22.02	14.046667	10.34265601	5.97134
215	18.38	2.2	22.63	14.403333	10.77991806	6.22379
217	15.51	4.78	20.17	13.486667	7.891985386	4.55644
219	17.93	4.64	20.36	14.31	8.462145118	4.88562
238	11.01	6.64		8.825	3.090056634	1.78405
240	9.81	4.21	19.48	11.166667	7.724871088	4.45996
242	11.53	3.73	15.76	10.34	6.102646967	3.52336
1992 Aug. 1	14.61	2.22	23.46	13.43	10.66905338	6.15978
3	14.02	2.02	21.53	12.523333	9.840733374	5.68155
5	13.59	1.79	20.34	11.906667	9.388867521	5.42067
7	15.68	2.18	23.89	13.916667	10.96189004	6.32885
9	20.98	3.51	36.57	20.353333	16.53890665	9.54874
11	23.26	7.44	30.12	20.273333	11.63123954	6.7153
19	16.23	8.53	25.69	16.816667	8.595029571	4.96234
Observed mear	12.8548148	2.49703704	18.2796154	11.093765		
Standard dev.	5.08881985	2.3246502	7.6459843	4.6514696		
Maximum	23.26	8.53	36.57	20.353333		
Median	13.59	2.18	19.61	11.166667		

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CH4 flux (mg/m2/day)						
Julian day	T1	T2	T3	Mean T	Standard dev.	St. error
166	4.13	0.74	2.57	2.48	1.696791089	0.97964
168	4.21	1.17	4.28	3.22	1.775697046	1.0252
170	5.18	3.12	6.6	4.9666667	1.749780939	1.01024
172	6.43	8.46	12.3	9.0633333	2.981146312	1.72117
175	14.27	19.72	20.45	18.146667	3.377074671	1.94975
180	31	25.01	13.87	23.293333	8.69306812	5.01895
187	26.62	25.74	13.04	21.8	7.599131529	4.38736
189	29.5	28.17	8.53	22.066667	11.74194334	6.77921
191	32.36	24.18	9.14	21.893333	11.77767945	6.79985
193	27.86	20.29	8.71	18.953333	9.64472049	5.56838
195	34.2	20.56	11.6	22.12	11.38047451	6.57052
205	77.09	34.61	26.66	46.12	27.11376588	15.6541
212	129.68	42.45	38.66	70.263333	51.49122482	29.7285
213	146.47	41.69	38.51	75.556667	61.43332755	35.4685
215	156.02	42.75	42.93	80.566667	65.34456545	37.7267
217	158.1	36.1	43.28	79.16	68.45824129	39.5244
219	187.76	36.46	51.13	91.783333	83.44125259	48.1748
238	193.06	52.08	67.03	104.05667	77.44075564	44.7104
240	162.07	43.5	67.64	91.07	62.66127113	36.1775
242	157.16	40.32	76.63	91.37	59.7983871	34.5246
1992 Aug. 1	70.9	56.34	13.02	46.753333	30.10733687	17.3825
3	82.72	54.93	21.78	53.143333	30.50926144	17.6145
5	101.01	50.1	25.4	58.836667	38.55470183	22.2596
7	130.04	58.45	34.53	74.34	49.69818005	28.6933
9	168.34	62.14	43.68	91.386667	67.279674	38.8439
11	183.12	65.63	49.68	99.476667	72.87493419	42.0744
19	252.41	54.41	114.73	140.51667	101.4875073	58.5938
Observed mear	95.2485185	35.1525926	32.0881481	54.163086		
Standard dev.	74.031134	18.8965505	26.7079039	37.973462		
Maximum	252.41	65.63	114.73	140.51667		
Median	82.72	36.46	25.4	53.143333		

Day Aug 1992	T4	T5	T6	T7		
7	4.52	0	0	-0.75		
9	6.5	1.03	0	0		
11	5.23	0.84	0.95	0		
19	5.47	0	0	-0.45		
21	5.54	0	0	-0.4		
Observed mear	5.452	0.374	0.19	-0.32		
Stdev	0.71138597	0.5165075	0.42485292	0.3213254		
Max	6.5	1.03	0.95	0		
Median	5.47	0	0	-0.4		

Appendix 3. Model results

The following are examples of outputs from the model experiments described in Chapter 6. Page 175 shows varies parameters from a stand-alone run with the model as described under experiment 1 in Chapter 6. The x-axis shows the julian day 1991. Pages 176-178 show results from a five year single column run with the model as described under experiment 3 in Chapter 6. TSL is temperature of soil layer. THETA is the unfrozen soil moisture. THETAS is the frozen soil moisture. Figures in brackets are means of the parameters in a given run.

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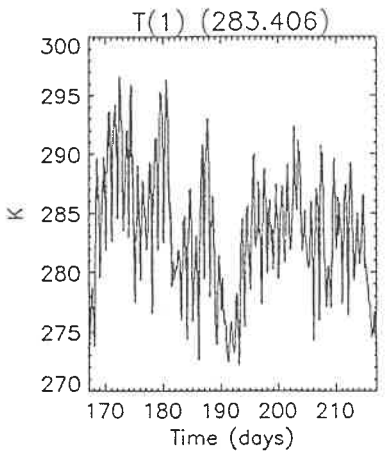
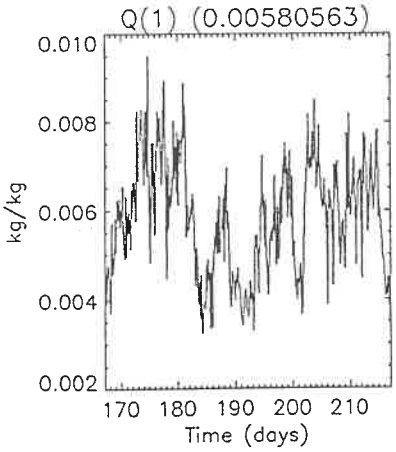
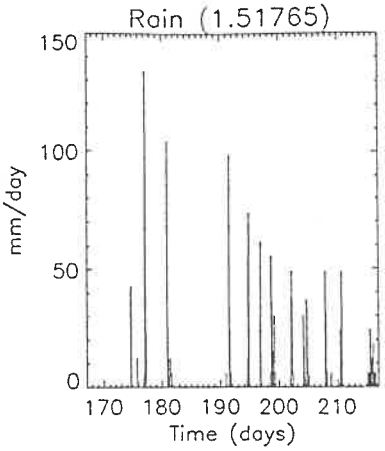
θ = 0.5, 0.6, 1.0, 1.0 α = 0.85

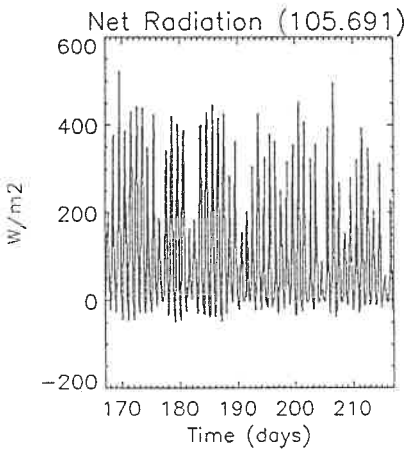
T = 285, 277, 273, 265 β = 0.00075

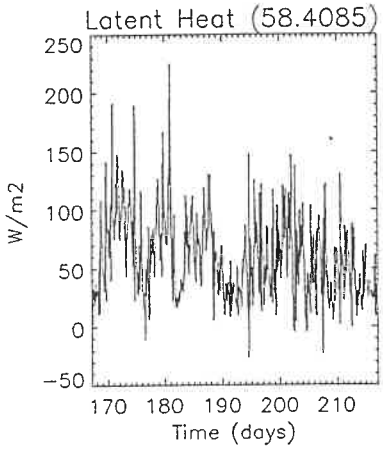
$\Delta T = 0$; $f_\theta = 1.0$

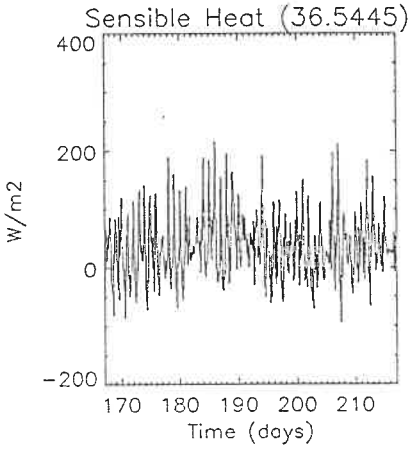


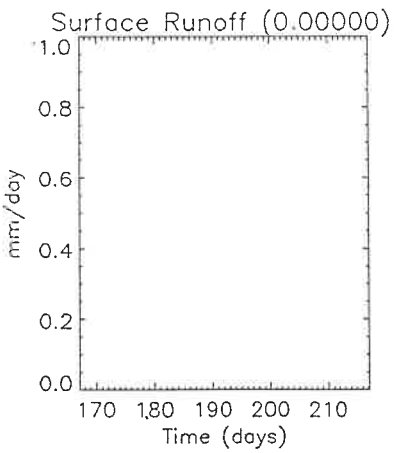


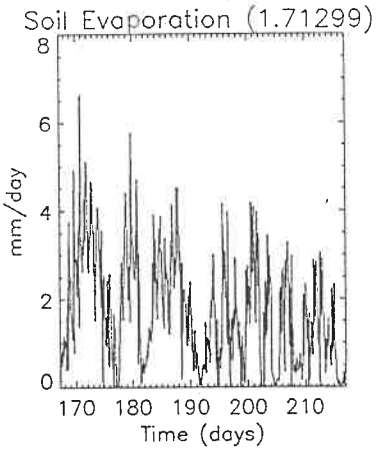


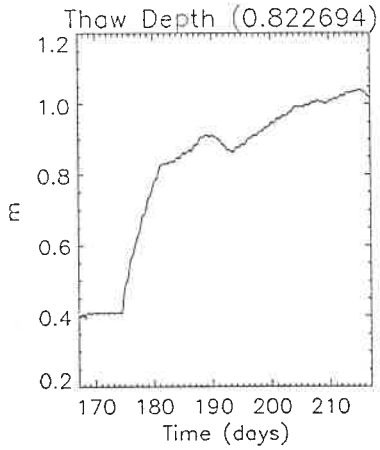


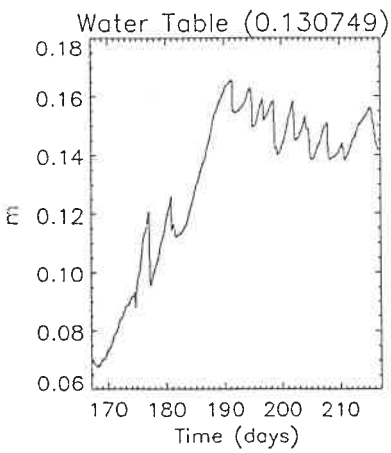


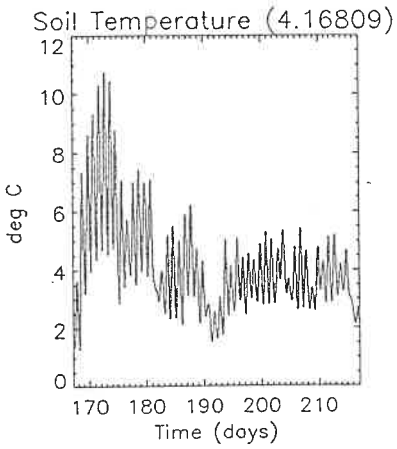


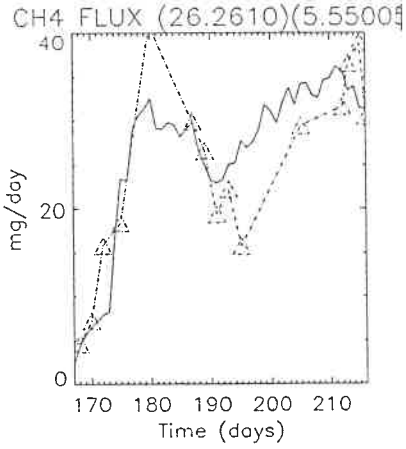




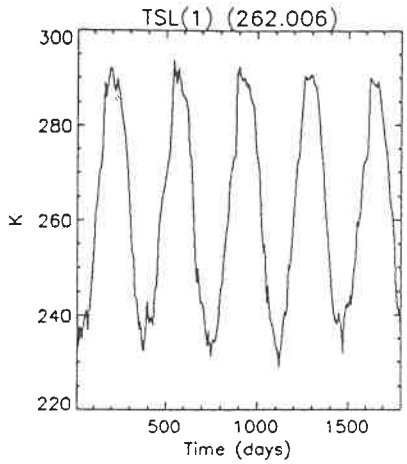


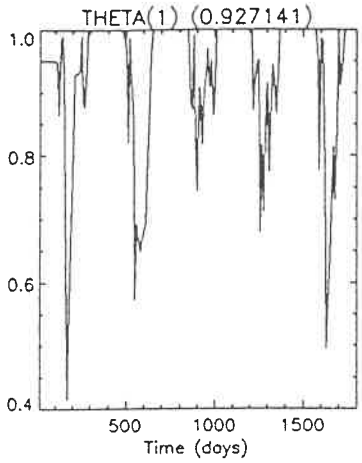


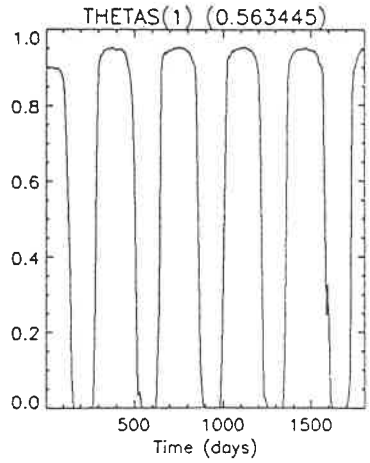


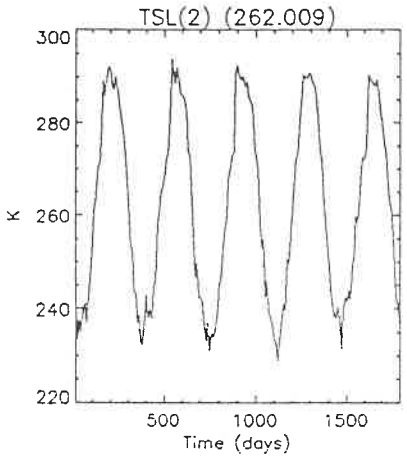


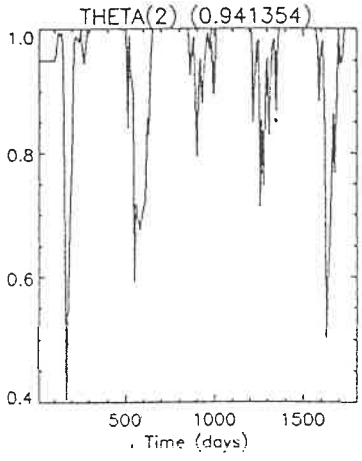
24/3/93 1xCO2 Five yr

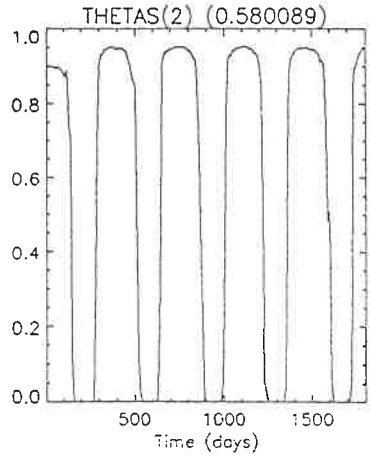


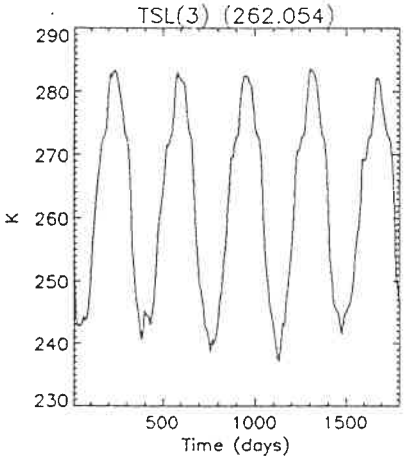


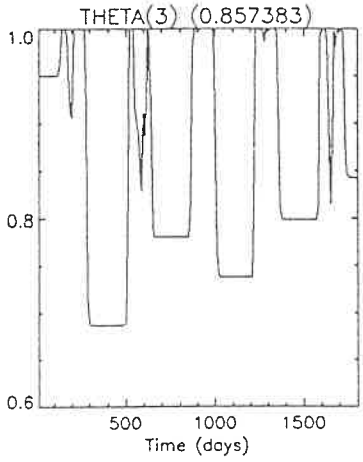


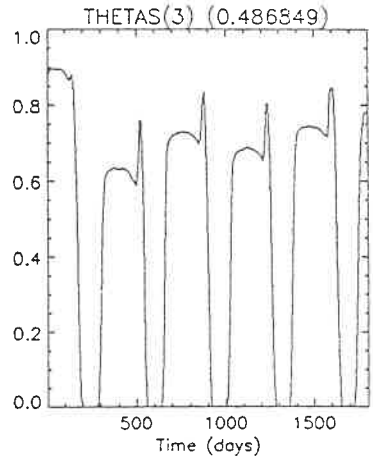


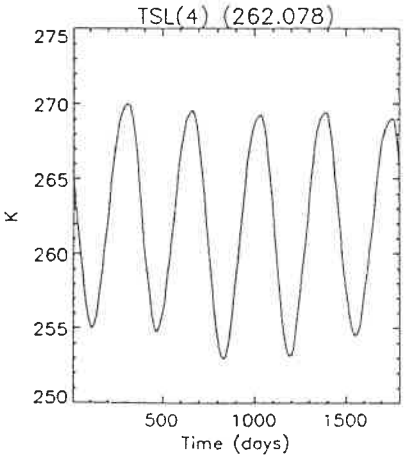


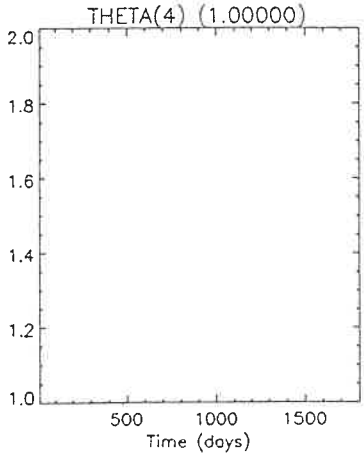


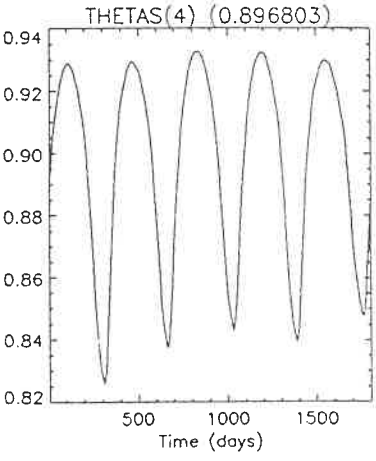


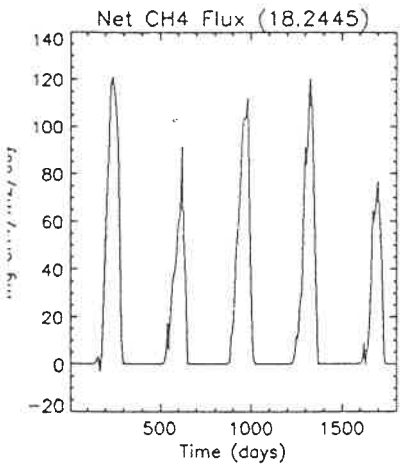








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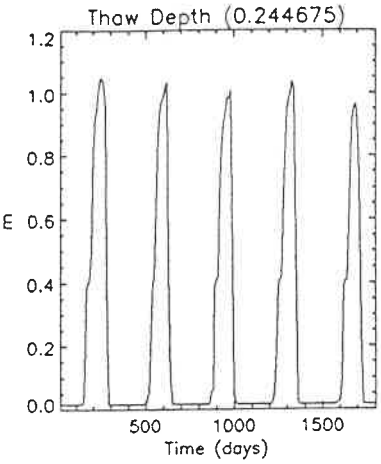
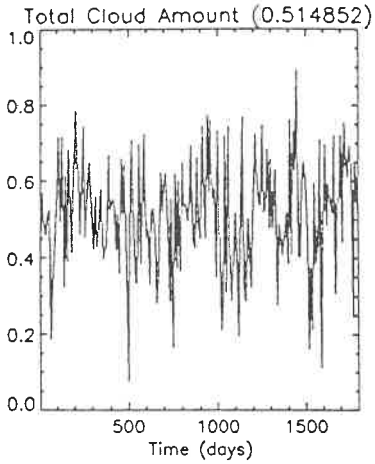
Time (days)


Thaw Depth (0.244675)
m
Time (days)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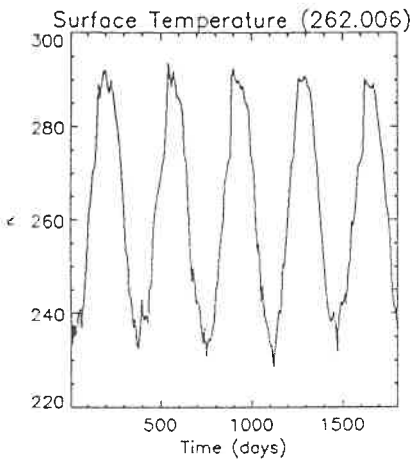
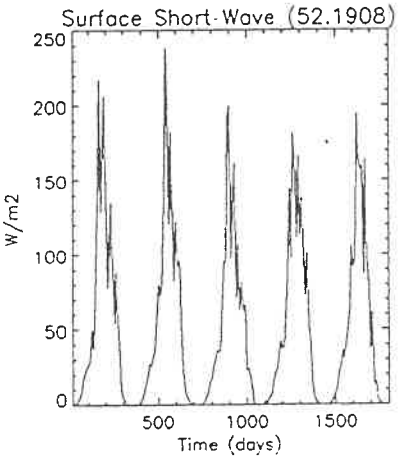
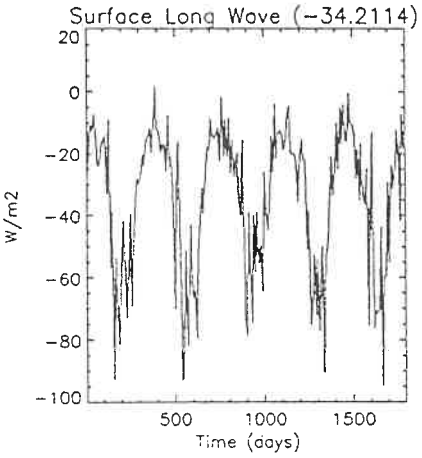
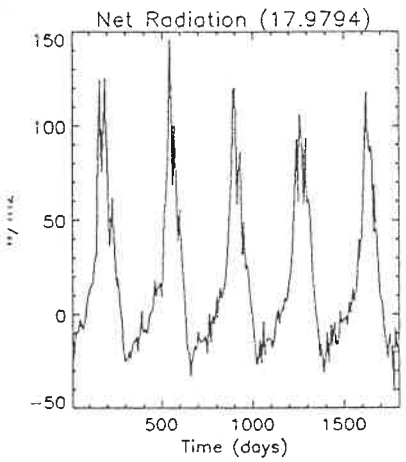
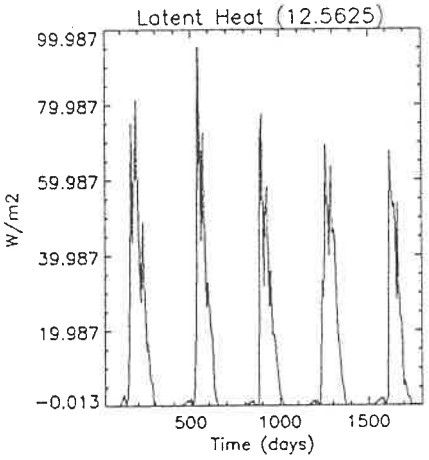
Time (days)

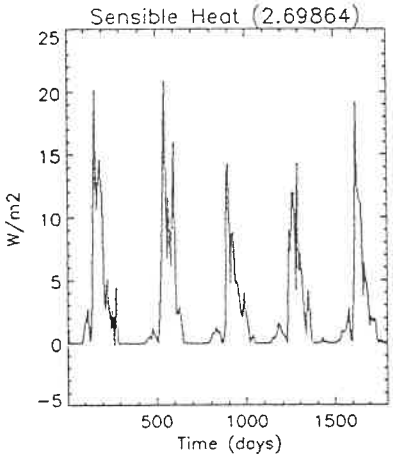

Surface Temperature (262.006)
K
Time (days)

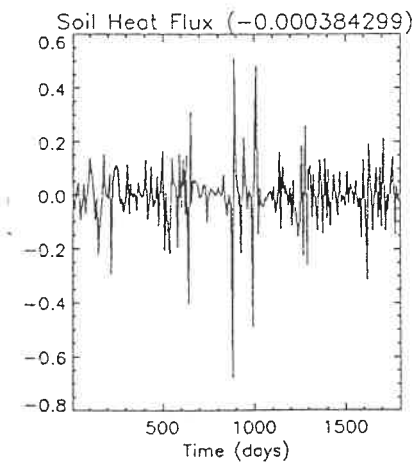

Surface Short-Wave (52.1908)
W/m2
Time (days)

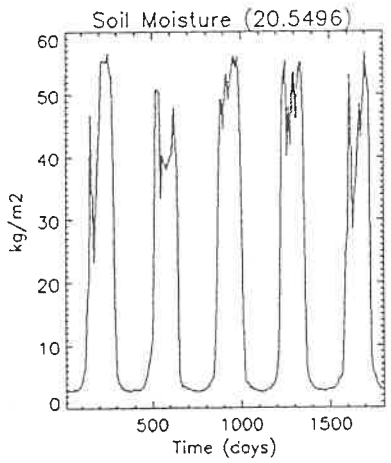

Surface Long Wave (-34.2114)
W/m2
Time (days)

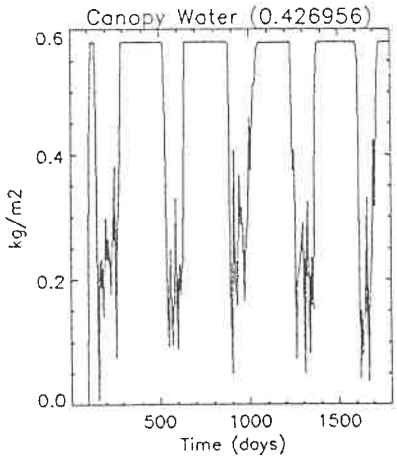

Net Radiation (17.9794)
W/m2
Time (days)


Latent Heat (12.5625)
W/m2
Time (days)


Sensible Heat (2.69864)
W/m2
Time (days)


Soil Heat Flux (-0.000384299)
Time (days)


Soil Moisture (20.5496)
kg/m2
Time (days)


Canopy Water (0.426956)
kg/m2
Time (days)